YAYASAN PERGURUAN CIKINI
INSTITUT SAINS DAN TEKNOLOGI NASIONAL
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SURAT PENUGASAN TENAGA PENDIDIK

Nomor : 16 /03.1-F/IV/2023
SEMESTER GENAP TAHUN AKADEMIK 2022/2023

Nama	:	**Muhamad Komarudin, S.Si., M.Si.**	**Status Pegawai**	:	**Tetap**
NIK	:	**01.221571**	**Program Studi**	:	**Teknik Sipil**
Jabatan Akademik	:	**NA**			

Bidang	Perincian Kegiatan	Tempat	Hari	Kredit (sks)	Ket
I PENDIDIKAN DAN PENGAJARAN	**1. MENGAJAR DI KELAS (KULIAH/RESPONSI DAN LABORATORIUM)**				
	1. Drainase & sanitasi lingkungan	S1 - Reg	Kamis	1	
	2. Ilmu ukur tanah	S1 - Reg	Rabu	1	
	3. Rekayasa Lingkungan	S1 - Reg	Kamis	1	
	4. Rekayasa Lingkungan	S1 - K	Selasa	1	
	5. Drainase & sanitasi lingkungan	S1 - K	Selasa	1	
	6. Ilmu ukur tanah	S1 - K	Kamis	1	
	7. Ilmu ukur tanah	D3 - Reg	Kamis	1	
	8. Praktikum Ukur Tanah & SIG	S1 - Reg		1	
	Penugasan sebagai Ka. Lab. Hidrologi dan SIG			3	
	2. PEMBIMBING				
	1. Seminar				
	2. Kerja Praktek				
	3. Tugas Akhir			1	
	4. Pembimbing Akademik			1	
	3. PENGUJI				
	1. Tugas Akhir				
	2. Kerja Praktek				
II PENELITIAN	1. Penelitian Ilmiah				
	2. Penulisan Karya Ilmiah			1	
	3. Penulisan Diktat Kuliah				
	4. Menerjemahkan Buku Kuliah				
	5. Pengembangan Program Kuliah Kurikulum				
	6. Pengembangan Bahan Ajar				
III PENGABDIAN DAN MASYARAKAT	1. Menduduki Jabatan di Pemerintah				
	2. Pengembangan Hasil Pend & Penelitian untuk Pengab Masyarakat				
	3. Memberikan Penyuluhan, Peltihan, Penataran, Ceramah pada Masyarakat				
	4. Memberikan Pelayanan Kepada Masyarakat Umum			1	
	5. Menulis Karya Pengabdian Pada Masyarakat yang Tidak Dipublikasikan				
	1. Menjadi Anggota Panitia/Badan pada suatu Perguruan Tinggi				
	2. Menjadi Anggota Badan Lembaga Pemerintah				
	3. Menjadi Anggota Organisasi Profesi			1	
	4. Mewakili PT/Lembaga Pemerintah, Duduk dalam Panitia antar Lembaga				
	5. Menjadi Anggota Delegasi Nasional ke Pertemuan-pertemuan International				
	6. Berperan Serta Aktif dalam Pertemuan Ilmiah / Seminar				
	7. Anggota Dalam Tim Penilai Jabatan Dosen				
Jumlah Total				**16**	

Kepada yang bersangkutan akan diberikan gaji/honorarium sesuai dengan peraturan penggajian yang berlaku di Institut Sains dan Teknologi Nasional
Penugasan ini berlaku tanggal 20 Maret 2023 sampai dengan 31 Agustus 2023

Jakarta, April 2023

Tembusan :
1. Direktur Akademik - ISTN
2. Direktur Non Akademik - ISTN
2. Ka. Biro Sumber Daya Manusia - ISTN
3. Kepala Program Studi Teknik Sipil
4. Arsip

FAKULTAS TEKNIK ... PERENCANAAN ISTN

(Ir. Lely Mustika, M.T.)

Analisa Dimensi Saluran

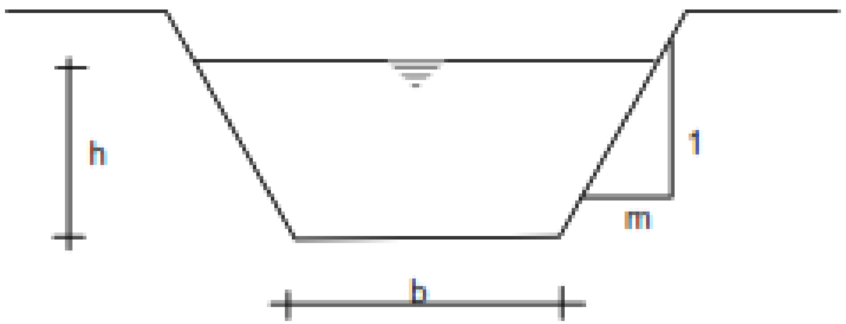
Muhamad Komarudin

PENAMPANG SALURAN EKONOMIS

Perencanaan saluran jaringan irigasi dan jaringan drainase yang pada dasarnya merupakan perencanaan penampang saluran terbuka yang mampu mengalirkan debit dari suatu lokasi ke lokasi lain dengan lancar, aman dan dengan biaya yang memadai. Dalam hal ini aspek-aspek yang perlu dipertimbangkan antara lain adalah :

- Aspek Hidrologis (Aspek Utama)
- Aspek ekonomi,
- Aspek keamanan lingkungan, dan
- Aspek estetika.

1. **Saluran Bentuk Trapesium**



Gambar 8 Saluran bentuk trapesium

Rumus yang digunakan :

$$A_e = (b + m.h)h$$

$$P = b + 2h\sqrt{(1 + m^2)}$$

$$R = \frac{A_e}{P}$$

Dimana :

B = lebar saluran (m)

h = dalamnya air (m)

m = perbandingan kemiringan talud

R = jari – jari hidrolis (m)

P = Keliling basah saluran (m)

A_e = Luas Penampang basah (m^2)

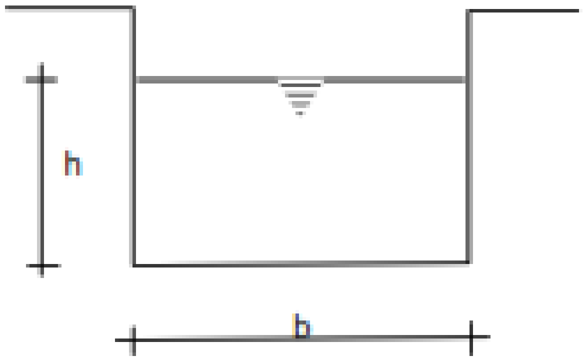
2. **Saluran Bentuk Segi Empat**

Rumus yang digunakan :

$$A_e = b.h$$

$$R = \frac{A_e}{P}$$

$$P = b + 2h$$



Gambar 9 Saluran bentuk segiempat

Dimana :

B = lebar saluran (m)

h = dalamnya air (m)

R = jari – jari hidrolis (m)

P = Keliling basah saluran (m)

A_e = Luas Penampang basah (m^2)

A.7.2 Penampang basah berdasarkan debit air (Q) dan kecepatan (V)

Dimensi saluran diperhitungkan dengan rumus Manning sebagai berikut :

$$Q = V.A$$

$$V = \frac{1}{n}(R)^{2/3}(i)^{1/2}$$

Dimana :

Q	:	Debit air di saluran (m^3/det)
V	:	Kecepatan air dalam saluran (m/det)
n	:	Koefisien kekasaran dinding.
R	:	Jari-jari hidraulik (meter)
i	:	Kemiringan dasar saluran
A	:	Luas penampang basah (m^2)

Tabel 20 Koefisien kekasaran dinding (n)

Tipe saluran	n
Lapisan beton	0,017 – 0,029
Pasangan batukali diplester	0,020 – 0,025
Saluran dari alam	0,025 – 0,045

A.7.3 Kemiringan Talud.

1. **Kemiringan Talud Saluran Tanah.**

   Kemiringan talud disesuaikan dengan karakteristik tanah setempat yang pada umumnya berkisar antara 1 : 1,5 s/d 1 : 4.

Tabel 21 Kemiringan Talud Bahan dari Tanah

Bahan Tanah	Kemiringan Talud (m = H/V)
Batu	0,25
Lempung kenyal, geluh	1 - 2
Lempung pasir, tanah kohesi f	1,5 - 2,5
Pasir lanauan	2 - 5
Gambut kenyal	1 - 2
Gambut lunak	3 - 4
Tanah dipadatkan dengan baik	1 - 1,5

2. **Kemiringan Talud Saluran Pasangan.**

Tabel 22 Kemiringan Talud Bahan dari Pasangan

Tinggi Air	m
h < 0,40 m	0 (dinding tegak vertikal)
0,75 > h > 0,40 m	0,25 - 0,5
H > 0,75 m	0,50 - 1,0

A.7.4 Tinggi Jagaan (F).

Tinggi jagaan minimum untuk saluran dengan pasangan direncanakan = 0,50m. Untuk saluran tanpa pasangan dengan debit tinggi jagaan sebagai berikut :

Tabel 23 Tinggi jagaan

Q	F (m)	Polder (m)
$Q < 5\ m^3/det$	0,20 – 0,30	0,75 – 1,00
$10\ m^3/det > Q > 5\ m^3/det$	0,30 – 0,50	1,00 – 1,25
$Q > 10\ m^3/det$	0,70 – 1,00	1,25 – 1,50

A.7.5 Kemiringan Tanah

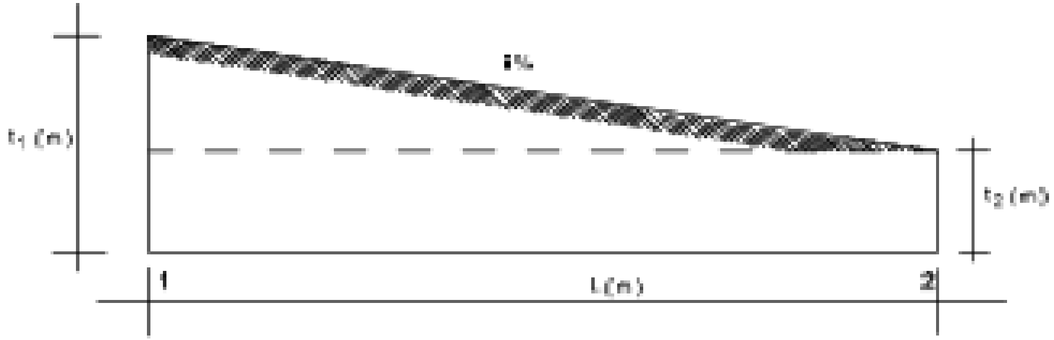
Kemiringan tanah di tempat dibuatnya fasilitas saluran drainase ditentukan dari hasil pengukuran di lapangan, dihitung dengan rumus :

$$i = \frac{t_1 - t_2}{L} x\, 100\%$$

Keterangan :

t_1 = tinggi tanah di bagian tertinggi (m)

t_2 = tinggi tanah di bagian terendah (m)



Gambar 10 Kemiringan tanah

Tabel 24 Harga n untuk rumus Manning

No	Tipe Saluran	Baik sekali	Baik	Sedang	Jelek
	SALURAN BUATAN				
1	saluran tanah, lurus teratur	0.017	0.02	0.023	0.025
2	saluran tanah yang dibuat dengan excavator	0.023	0.028	0.03	0.04
3	saluran pada dinding batuan, lurus, teratur	0.02	0.03	0.033	0.035
4	saluran pada dinding batuan, tidak lurus, tidak teratur	0.035	0.04	0.045	0.045
5	saluran batuan yang diledakkan, ada tumbuh-tumbuhan	0.025	0.03	0.035	0.04
6	dasar saluran dari tanah, sisi saluran berbatu	0.028	0.03	0.033	0.035
7	saluran lengkung, dengan kecepatan aliran rendah	0.02	0.025	0.028	0.03
	SALURAN ALAM				
8	Bersih, lurus tidak berpasir, tidak berlubang	0.025	0.028	0.03	0.033
9	seperti no.8, tetapi tidak ada timbunan atau kerikil	0.03	0.033	0.035	0.04
10	Melengkung bersih, berlubang dan berdinding pasir	0.033	0.035	0.04	0.045

Debit air yang masuk :

$$\begin{aligned} Q_{in} &= 0{,}278C \times C_s \times I \times A \\ &= 0{,}278 \times 0{,}73 \times 0{,}70 \times 59 \times 5 \\ &= 42 m^3 / \text{det} \end{aligned}$$

Contoh Perhitungan 7 :

Analisa dimensi saluran trapesium dengan menggunakan data perencanaan sebagai berikut

- Debit air yang masuk (Q_{in}) = 42 m^3/det (diambil dari contoh perhitungan 5)
- Lebar saluran (b) = 5 m
- Dalamnya air (h) = 1,9 m
- Perbandingan kemiringan talud (m) = 1,5
- Kemiringan saluran yang diijinkan (i) = 0,0025
- Koefisien kekasaran Manning (n) = 0,020

Gambar 11 Kemiringan tanah

Penyelesaian :

1) Luas penampang basah saluran :

$$\begin{aligned} A_e &= (b + m.h)h \\ &= (5{,}0 + (1{,}5x1{,}9))x1{,}9 \\ &= 14{,}92 m^2 \end{aligned}$$

2) Keliling basah saluran :

$$\begin{aligned} P &= b + 2h\sqrt{(1 + m^2)} \\ &= 5 + 2(1{,}9)\sqrt{(1 + (1{,}5)^2)} \\ &= 11{,}9 m \end{aligned}$$

3) Jari-jari hidrolis :

$$\begin{aligned} R &= \frac{A_e}{P} \\ &= \frac{14{,}92}{11{,}9} \\ &= 1{,}26 m \end{aligned}$$

4) Kecepatan aliran :

$$\begin{aligned} V &= \frac{1}{n}(R)^{2/3}(i)^{1/2} \\ &= \frac{1}{0{,}020}(1{,}26)^{2/3}(0{,}0025)^{1/2} \\ &= 2{,}91 m / \text{det} \end{aligned}$$

5) Debit air yang keluar :

$$\begin{aligned} Q_{out} &= V.A \\ &= 2{,}91 \times 14{,}92 \\ &= 43{,}47 m^3 / \text{det} \end{aligned}$$

6) Check :

$$\begin{aligned} R_{cm} &= \frac{Q_{in}}{Q_{out}} \\ &= \frac{42}{43{,}47} \\ &= 0{,}97 \qquad (OK) \end{aligned}$$

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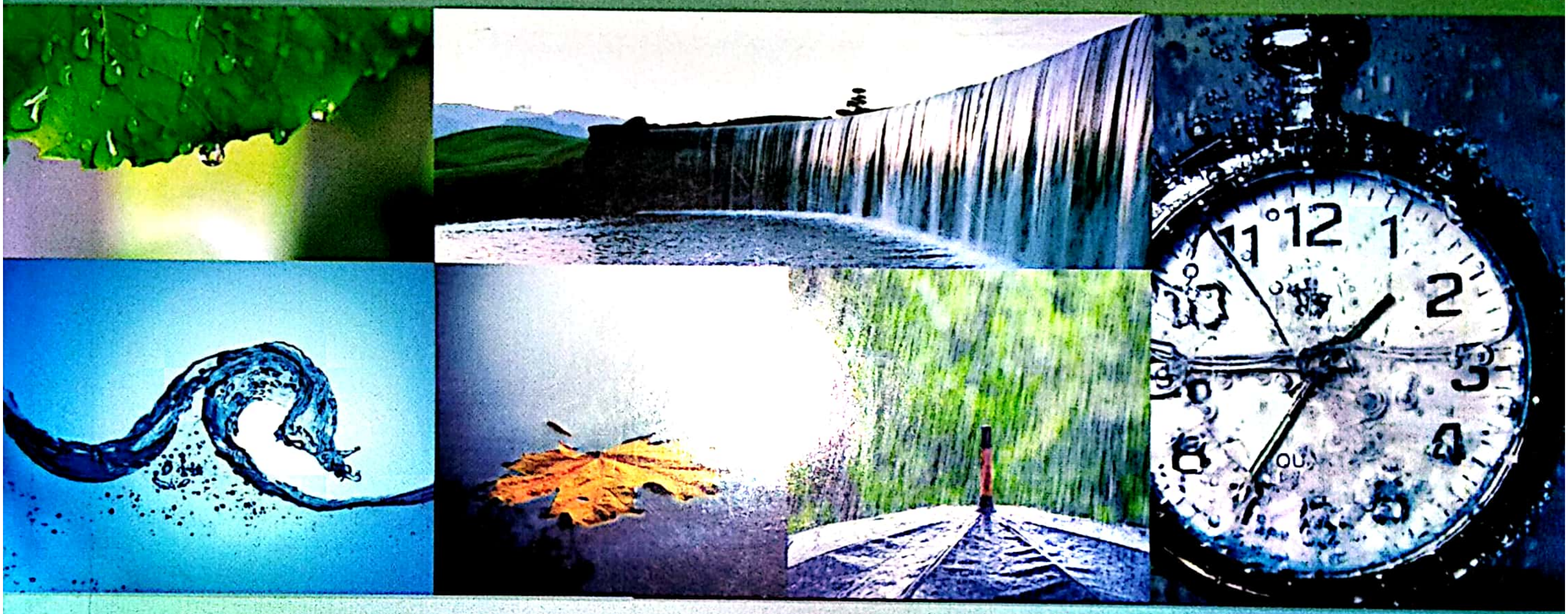

HIDROLOGI TERAPAN

Dr. Ir. A. Syarifudin, M.Sc., PU-SDA

HIDROLOGI TERAPAN
Oleh: Dr. Ir. A. Syarifudin, M.Sc., PU-SDA

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KATA PENGANTAR

Puji syukur ke hadirat Allah SWT, Rabb semesta alam dan penguasa hari pembalasan, Shalawat dan Salam senantiasa tercurah kepada Rasulullah SAW, atas taufiq dan hidayah-Nya sehingga penulis dapat menyelesaikan Buku "**HIDROLOGI TERAPAN**".

Dalam buku ini diuraikan secara singkat tentang pengertian hidrologi, siklus hidrologi dan sejarah perkembangan hidrologi (Bab 1), Jaringan Sungai dan Topografi (Bab 2), Kriteria Perhitungan Debit Banjir (Bab 3), Penelusuran Banjir (Bab 4), dan Metode Perhitungan Debit Banjir (Bab 5).

Buku ini sengaja memasukkan studi kasus penanganan sungai Lempuing yang ada kaitannya dengan analisis hidrologi, agar pembaca nantinya dapat mengetahui konsep dalam penerapan analisis hidrologi di lapangan.

DAFTAR ISI

HIDROLOGI TERAPAN

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Untuk Istri & Anak2:

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Henggar Risa Destania, ST, M.Eng

Adie Yudha Prawira, ST

Naufallah Dinda Harumi

SAMBUTAN

Sebelum saya dipercaya oleh Menteri Pekerjaan Umum dan Perumahan Rakyat Republik Indonesia sebagai Direktur Jenderal Bina Marga, saya banyak berkecimpung di Bidang Sumber Daya Air jadi saya tahu persis bahwa di Bidang keairan sangat diperlukan buku-buku yang berkaitan dengan SDA dan Bangunan Air.

Dengan diterbitkannya buku di bidang Teknik Sipil khususnya di bidang Sumber Daya Air (SDA) perlu disambut baik karena hal tersebut berarti semakin menambah perbendaharaan buku terutama diperuntukkan bagi mahasiswa.

Semoga buku ini bermanfaat bagi pembaca dan mahasiswa S-1 dalam menganalisis data curah hujan menjadi debit rancangan untuk bangunan keairan.

Jakarta,   April 2017

Direktur Jenderal Bina Marga
Kementerian PUPR,

Dr. Ir. Arie Setiadi Moerwanto, M.Sc

KATA PENGANTAR

Puji syukur ke hadirat Allah SWT Rabb semesta alam dan penguasa hari pembalasan, Shalawat dan Salam senantiasa tercurah kepada Rasulullah SAW, atas taufiq dan hidayah-Nya sehingga penulis dapat mentelesaikan Buku "HIDROLOGI TERAPAN"

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Penulis menyadari betul bahwa buku ini masih banyak terdapat kekurangan, oleh sebab itu dengan kerendahan hati penulis sangat mengharapkan masukan dan saran yang dapat menyempurnakan buku ini, sehingga dapat bermanfaat bagi kita semua, Amin

Palembang, Februari 2017

Penulis,

DAFTAR ISI

DAFTAR GAMBAR

DAFTAR TABEL

1 PENDAHULUAN

A. Pengertian Hidrologi

adalah sirkulasi air yang tidak pernah berhenti dari atmosfir ke bumi dan kembali ke atmosfir melalui kondensasi, presipitasi, evaporasi dan transpirasi.

Presipitasi

Presipitasi pada pembentukan hujan, salju dan hujan batu (*hail*) yang berasal dari kumpulan awan. Awan-awan tersebut bergerak mengelilingi dunia, yang diatur oleh arus udara. Sebagai contoh, ketika awan-awan tersebut bergerak menuju pegunungan, awan-awan tersebut menjadi dingin, dan kemudian segera menjadi jenuh air yang kemudian air tersebut

Kondensasi (pengembunan)

Ketika uap air mengembang, mendingin dan kemudian berkondensasi, biasanya pada partikel-partikel debu kecil di udara. Ketika kondensasi terjadi dapat berubah menjadi cair kembali atau langsung berubah menjadi padat (es, salju, hujan batu (*hail*)). Partikel-partikel air ini kemudian berkumpul dan membentuk awan.

Evaporasi (penguapan)

Ketika air dipanaskan oleh sinar matahari, permukaan molekul-molekul air memiliki cukup energi untuk melepaskan ikatan molekul air tersebut dan kemudian terlepas dan mengembang sebagai uap air yang tidak terlihat di atmosfir.

Sekitar 95.000 mil kubik air menguap ke angkasa setiap tahunnya. Hampir 80.000 mil kubik menguapnya dari lautan. Hanya 15.000 mil kubik berasal dari daratan, danau, sungai, dan lahan yang basah, dan yang paling penting juga berasal dari tranpirasi oleh daun tanaman yang hidup. Proses semuanya itu disebut Evapotranspirasi.

Perkolasi

Beberapa presipitasi dan salju cair bergerak ke lapisan bawah tanah, mengalir secara infiltrasi atau perkolasi melalui celah-celah dan pori-pori tanah dan batuan sehingga mencapai muka air tanah (*water table*) yang kemudian menjadi air bawah tanah.

Pemanasan air samudera oleh sinar matahari merupakan kunci proses siklus hidrologi tersebut dapat berjalan secara kontinu. Air berevaporasi, kemudian jatuh sebagai presipitasi dalam bentuk hujan, salju, hujan batu, hujan es dan salju (*sleet*), hujan gerimis atau kabut.

Pada perjalanan menuju bumi beberapa presipitasi dapat berevaporasi kembali ke atas atau langsung jatuh yang kemudian diintersepsi oleh tanaman sebelum mencapai tanah. Setelah mencapai tanah, siklus hidrologi terus bergerak secara kontinu dalam tiga cara yang berbeda:

- Evaporasi / transpirasi - Air yang ada di laut, di daratan, di sungai, di tanaman, dsb. kemudian akan *menguap ke angkasa (atmosfer)* dan kemudian akan menjadi awan. Pada keadaan jenuh uap air (awan) itu akan menjadi bintik-bintik air yang selanjutnya akan turun (precipitation) dalam bentuk hujan, salju, es.

- Infiltrasi / Perkolasi ke dalam tanah - Air bergerak ke dalam tanah melalui celah-celah dan pori-pori tanah dan batuan menuju muka air tanah. Air dapat bergerak akibat aksi kapiler atau air dapat bergerak secara vertikal atau horizontal dibawah permukaan tanah hingga air tersebut memasuki kembali sistem air permukaan.
- Air Permukaan - Air bergerak diatas permukaan tanah dekat dengan aliran utama dan danau; makin landai lahan dan makin sedikit pori-pori tanah, maka aliran permukaan semakin besar. Aliran permukaan tanah dapat dilihat biasanya pada daerah urban. Sungai-sungai bergabung satu sama lain dan membentuk sungai utama yang membawa seluruh air permukaan disekitar daerah aliran sungai menuju laut.

Air permukaan, baik yang mengalir maupun yang tergenang (danau, waduk, rawa), dan sebagian air bawah permukaan akan terkumpul dan mengalir membentuk sungai dan berakhir ke laut. Proses perjalanan air di daratan itu terjadi dalam komponen-komponen siklus hidrologi yang membentuk sisten Daerah Aliran Sungai (DAS).*Jumlah air di bumi* secara keseluruhan relatif tetap, yang berubah adalah wujud dan tempatnya.

Secara umum dapat dikatakan bahwa hidrologi adalah ilmu yang menyangkut masalah kuantitas dan kualitas air di bumi.

Hidrologi dapat dikatagorikan menjadi 2 (dua) bagian :

- *Operational Hydrology*

  Menyangkut pemasangan alat-alat ukur berikut penentuan jaringan stasiun pengamatannya, pengumpulan data hidrologi (pengamatan elemen-elemen hidrologi), pengolahan data mentah dan publikasi data.

- *Applied Hydrology*

  Ilmu yang langsung berhubungan dengan penggunaan hukum-hukum yang berlaku menurut ilmu-ilmu murni pada kejadian praktis dalam kehidupan menyangkut analisa hidrologi.

  Contoh :

Pada kegiatan perencanaan reservoir yang bertujuan untuk mengendalikan banjir dan mengatasi kebutuhan air, tercakup beberapa langkah analisa hidrologi adalah :

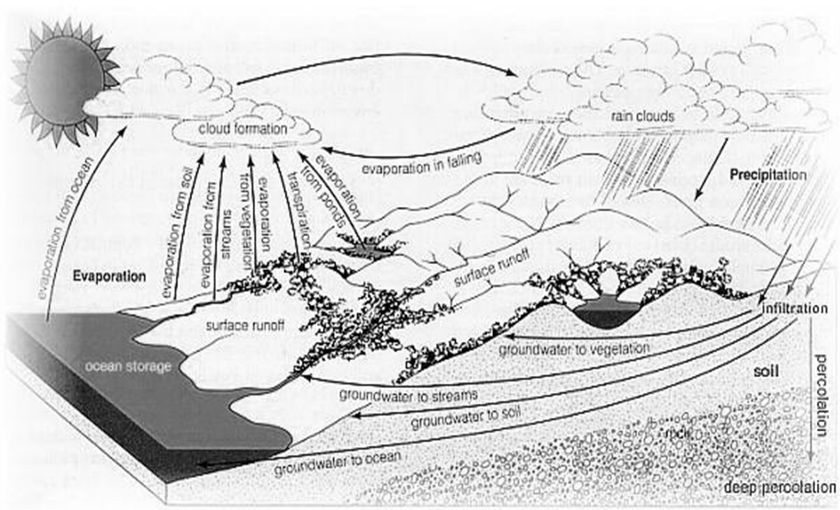
- ✓ Memperkirakan jumlah air permukaan yang tersedia.
- ✓ Memperkirakan kehilangan air (akibat penguapan, rembesan dsb)
- ✓ Memperkirakan kebutuhan air (domistik, pertanian, perindustrian).
- ✓ Memperkirakan banjir rencana/design flood.
- ✓ Memperkirakan kapasitas/volume reservoir dan tinggi muka air maksimum dalam reservoir.

Setelah itu baru dilanjutkan dengan perencanaan bangunan air yaitu :

- ✓ Merencanakan bangunan pengendalian banjir.

- ✓ Merencanakan bangunan drainase pada daerah perkotaan atau daerah aliran.
- ✓ Merencanakan / menentukan bentuk dan ukuran konstruksi dll.

B. Siklus Hidrologi



sumber:

Gambar 1.1. Skema siklus hidrologi

Akibat panas yang bersumber dari sinar matahari, maka akan terjadi :

- Evaporasi : penguapan pada permukaan air sungai, danau, waduk dan pada permukaan tanah.
- Transpirasi : penguapan dari permukaan tanaman.

Uap air hasil penguapan ini pada ketinggian tertentu akan menjadi awan, kemudian karena beberapa sebab awan akan berkondensasi menjadi presipitasi (bisa dalam bentuk salju, hujan es, hujan, embun).

Air hujan yang jatuh kadang-kadang tertahan oleh ujung daun atau oleh bangunan dan sebagainya. Hal ini diberi istilah intersepsi, dimana besarnya intersepsi pada tanaman tergantung dari jenis tanaman, tingkat pertumbuhan, tetapi biasanya berkisar 1 mm pada hujan-hujan pertama, kemudian sekitar 20% pada hujan-hujan berikutnya.

Air hujan yang mencapai tanah sebagian terinfiltrasi (menembus permukaan tanah), sebagian lagi menjadi aliran air di atas permukaan (over-land flow) kemudian terkumpul di slauran. Aliran ini disebut *surface run-off.*

Hasil infiltrasi sebagian mengalir menjadi aliran air bawah permukaan (inter-flow/sub-surface flow/through flow), sebagian lagi akan membasahi tanah.

Air yang menjadi bagian dari tahah dan berada dalam pori-pori tanah disebut *air soil.*

Apabila kapasitas kebasahan tanah / soil moisture ini terlampaui, maka kelebihan airnya akan ber perkolasi (mengalir vertikal) menjadi air tanah.

Aliran air tanah (ground water flow) akan terjadi sesuai dengan hukum-hukum fisika.

Air yang mengalir itu pada suatu situasi dan kondisi tertentu akan mencapai danau, sungai dan laut menjadi depression storage (simpanan air yang disebabkan oleh cekungan), saluran dan sebagainya, mencari tempat lebih rendah.

Sehingga secara garis gesar, pada sistem sirkulasi ini dapat dikatagorikan veriabel-veriabel mana yang berperan sebagai input dan mana yang berperan sebagai output.

C. Sejarah pengembangan hidrologi

1. Pengembangan awal

Aplikasi praktis dalam hidrologi telah mulai diterapkan, misalkan pada pembuatan:

- Sumur-sumur zaman purba di arab.

- Sistem irigasi di cina
- Dam/reservoir air terbesar di dunia pada saat itu (± 4800 tahun yang lalu di Mesir).

Design hidrologi dari pekerjaan teknik hidraulik hanya didasarkan pada pengalaman dan pengamatan.

2. Pengembangan pada abad sesudah masehi

Secara praktis, ilmu hidrologi baru dimulai pada abad 16 sejak :

- Leonardo da Vinci & Bernard Palissy menemukan konsep siklus hidrologi secara benar, melalui penyelidikan (hubungan infiltrasi sampai kepada terjadinya mata air).
- Pierre Perrault & Edme Mariotte (1686) mengadakan pengukuran aliran sungai pertama kali (pengukuran penampang melintang & kecepatan aliran), kemudian membandingkannya dengan hujan dan evaporasi DAS sehingga dengan adanya alat pengukur & pengembangan hidraulika, membuka kemungkinan dilaksanakan percobaan-percobaan hidrologi.

3. Hidrologi modern

- 1850 – 1900 : pengukuran-pengukuran sesaat dari debit
- 1900 – 1930 : periode penggunaan rumus-rumus empiris (mulai dilakukan pengumpulan data muka air – debit sungai secara sistematis).
- 1930 – 1950 : periode penggunaan konsep secara rational (teori infiltrasi, teori unit hidrograf, pengembangan hidraulika air tanah, rumus-rumus semi empiris).
- 1950 – sekarang : periode penggunaan teori-teori (analisa linier & non linier dari sistem hidrologi, teori unstedy flow dalam air tanah, aplikasi & teori mass-transfer menjadi analisa evaporasi, studi dari dinamika soil moisture, pengumpulan dari data hidrologi

yang berkesinambungan).

- Terbaru : penggunaan alat-alat modern sinar gamma, sinar laser, super sonic & planet) untuk berbagai tujuan penyelidikan dan pengumpulan data dll.

2 JARINGAN SUNGAI & TOPOGRAFI

A. Jaringan Sungai

Sungai Lempuing adalah anak Sungai Komering atau sungai orde ketiga dari Sungai Musi. Bentuk daerah pengaliran Sungai Lempuing secara umum berbentuk bulu burung dan sungai sejajar luas Catchment Area Sungai Lempuing sekitar sekitar 2.800 km^2.

Topografi DAS Lempuing dapat di bagi 3, hulu, tengah dan hilir:

a. Daerah hulu bergelombang tidak begitu tinggi antara 2 – 10 m dengan luas 1.247,00 km^2 (45%).

b. Daerah tengah relatif datar, alur-alur aliran sungai masih dapat terlihat dan berbentuk seluas 422,53 km^2 (15%).

c. Daerah hilir berupa rawa-rawa dan danau-danau seluas 270,53 km^2 (10%).

Daerah pengaliran sungai bagian hulu berupa pengaliran sejajar, terdiri dari 3 anak sungai Way Hitam, S. Belitang dan S.

Macak. Daerah pengaliran sedemikian mempunyai debit banjir yang kecil pada masing-masing sungai, oleh karena tibanya banjir dari masing-masing anak sungai berbeda-beda. Sebaliknya banjirnya berlangsung lama. Sungai-sungai sejajar tersebut mengumpul pada bagian hilir terjadi penjumlahan debit yang besar.

DAS bagian tengah dimulai dari pertemuan S. Macak sampai S. Burnai dan beberapa anak sungai lainya, merupakan lokasi banjir beberapa tahun belakangan ini, karena perubahan lahan akibat banyak terbuka.

DAS bagian hilir dimulai dari Muara Burnai sampai pertemuan dengan Sungai Komering terdapat beberapa anak-anak sungai merupakan daerah rawa dan danau-danau yang terjadi akibat air luapan sungai yang tidak dapat mengalir secara grafitasi.

Topografi daerah pengaliran tengah sampai hilir, merupakan sungai aluvial atau sungai yang mengalir pada lahan yang dibentuknya sendiri. Pada bagian hilir pertemuan dengan Sungai Komering tampang basah sungai hampir tidak kelihatan, karena topografi lahan dataran rawa-rawa.

Penanganan banjir pada DAS Lempuing bagian hilir perlu kajian dengan mempertimbangkan pemanfaatan danau-danau tersebut sebagai retensi banjir.

Tabel 2.1. Sungai sungai pada Sub DAS Lempuing

No	Nama Sungai	Panjang (km)	DAS (km^2)	Slope	Keterangan
A	Sungai Musi	480.00	59,942.00	0.00003	Q_2 = 4.000 m^3/dt (PLB)
A.1	Sungai Komering	190.00	9,980.00	0.00018	Q_2 = 900 m^3/dt
A.1.1	Sungai Lempuing	127.00	2,800.00	0.00027	
1	S. Danau	15.00	112.50	0.00015	sungai daerah rawa
2	S. Kemodo	8.00	22.40	0.00023	sungai daerah rawa
3	S. Bwg Batuhampar	17.50	66.50	0.00023	sungai daerah rawa
4	S. Beringin	9.50	32.00	0.00023	sungai daerah rawa
5	S. Penyengat	10.50	37.13	0.00023	sungai daerah rawa
6	S. Burnai	35.00	314.00	0.00041	
7	S. Petalang	4.00	5.60	0.00030	
8	S. Kemalajimat	7.50	21.00	0.00028	
9	S. Sepungkut	4.50	7.65	0.00028	
10	S. Belidah	5.00	21.00	0.00028	
11	S. Durian	11.00	25.00	0.00030	
12	S. Kepahyang	2.00	2.10	0.00030	
13	S. Mandarjaman	2.50	3.20	0.00035	
14	S. Sejengkah	2.50	2.00	0.00035	
15	S. Tuan Penghulu	3.00	2.55	0.00035	
16	S. Deras	22.95	9.00	0.00035	
17	S. Ketara	0.80	9.62	0.00035	
18	S. Macak	55.00	409.00	0.00033	
19	S. Belitang	57.00	388.00	0.00065	
20	S. Way Hitam	38.00	450.00	0.00070	
21	Lain-laian				

Tabel 2.2. Prosentase Luas DAS Lempuing

Persentase Luas DAS (km^2)			Panjang (km)
Hilir	270.53	0.10	60.50
Tengah	422.72	0.15	100.75
Hulu	1,247.00	0.45	150.00
Tersebar	859.75	0.31	
Total Luas DAS	2,800.00	1.0	311.25

B.Topografi dan Kondisi Banjir

Daerah hulu dan tengah sebagai pensuplai banjir dan hilir sebagai penampung. Terjadinya palung sungai seperti kondisi yang ada merupakan petunjuk bahwa bagian hulu dan tengah tampang sungai masih terbentuk oleh fluktuasi aliran, sedangkan bagian hilir tidak jelas, karena telah berfungsi sebagai penampung banjir dalam waktu yang lama.

Kondisi S. Komering pada lokasi pertemuan dengan S. Lempuing, merupakan daerah dataran banjir yang lebar, setiap tahun pada musim hujan selalu tergenang akibat luapan sungai. Kondisi ini harus dikaji sebab musababnya.

Kondisi lahan hilir sedemikian kesulitan tentang drainase, karena kemiringan topografinya sangat landai. Bila tidak lahan akan kebanjiran, air pergi dalam waktu lama. Pemanfaatan lahan akan merubah prilaku hidrotopografi daerah ini. Perubahan hidrotopografi tersebut seperti daerah-daerah parkir air banjir hilang jadi sawah dan pemukiman.

C. Corak Daerah Pengaliran

Corak daerah pengaliran diperlihatkan oleh suatu koefisien corak (F) yang merupakan perbandingan luas dan kuadrat panjang sungai (Soeyono Sostrodarsono & Kensku TAKEDA, 1976).

$F = A/L^2$

F = koefisien corak

A = luas daerah pengaliran (km^2)

L = panjang sungai (km)

Semakin besar F, makin lebar daerah pengaliran, contoh sungai dengan DAS bentuk bujur sangkar F =1. Daerah pengaliran seperti bentuk radial atau bentuk kipas atau lingkaran termasuk nilai F nya besar. Dimana anak-anak sungai terkonsentrasi ke satu titik secara radial. Daerah pengaliran sedemikian mempunyai banjir yang besar didekat titik pertemuaan anak-anak sungai.

Tabel 2.3. Koefisien Corak Sungai

Nama Sungai	Daerah Pengaliran (1.000 km^2)	Panjang Sungai Utama (km)	Koefisien Corak (F)
1. Amazon	7.050	6.200	0,183
2. Mississipi	3.250	6.500	0,077
3. Yangtze	1.780	5.200	0,066
4. Donau	620	2.900	0,097
5. Kiso (tiga sungai)	9,1	229	0,175

F = 0,50 – 1,00 sangat lebar

F = 0,25 – 0,50 lebar

F = 0,10 – 0,25 medium

F = 0,00 – 0,10 sempit

Sungai Lempuing F = $2.800/127^2 = 0{,}173$, termasuk klasifikasi medium, sampainya hujan di sungai utama lambat, mencapai outlet hilir lama.

D. **Kerapatan Sungai**

Untuk menilai kemampuan mendrain suatu DAS tergantung juga pada kerapatan jaringan sungai disamping kelandaian topografi dan kondisi geologi DAS. Kerapatan sungai adalah suatu indeks yang menunjukan banyaknya anak sungai pada suatu daerah pengaliran.

$K_s = (L_{su} + L_{an})/A$

Dengan :

K_s = Kerapatan sungai

L_{su} = panjang sungai utama (km)

L_{an} = panjang seluruh anak-anak sungainya (km)

A = luas daerah pengaliran (km^2)

Biasanya harga ini adalah kira-kira 0,10 sampai 1,00 (Djohar ND, 2002) dan dianggap sebagai indeks yang menunjukan keadaan topografi dan geologi dalam daerah pengaliran. Kerapatan besar artinya sungai dan anak sungainya banyak / panjang kemampuan mendrain lebih baik, sebaliknya kecil sungai dan anak sungai sedikit / pendek kemampuan mendrain kurang baik.

Sungai Lempuing

L_{su} = 127,00 km

L_{an} = 300,25 km

A = 2.800 km^2

K_s = (127+ 300,25)/2.800 = 0,152 / km,

Sungai Lempuing termasuk ber K_s kecil, anak sungai kurang dan kemampuan mendrain kurang baik. Karena air hujan jatuh pada lahan memerlukan waktu lama mencapai daerah rendah terdekat.

3 KRITERIA DESAIN

Seperti dalam kasus pada DAS Lempuing yang ada di Kabupaten OKI Provinsi Sumatera Selatan tidak tersedia data pengukuran aliran yang cukup untuk dianalisa secara statistik. Direktorat Jenderal Pengairan tahun 1999 menerbitkan Panduan Perencanaan Bendungan Tipe Urugan, Volume II Analisis Hidrologi dan PSA-007 menyarankan untuk menggunakan cara pengalih ragamkan curah hujan menjadi limpasan permukaan melalui Metode Hidrograf Satuan Sintetik.

3.1. Rumusan Debit Banjir

Metode Rasional adalah rumus perhitungan debit paling tua di dunia, dengan anggapan yang sangat sederhana. *Debit adalah tinggi hujan per satuan waktu jatuh pada luasan tertentu*.

$$Qp = C.i.A \qquad \text{..} \; (3.1)$$

Qp adalah debit puncak (satuan volume per waktu), *i* adalah intensitas hujan (tinggi hujan per waktu), C koefisien yang tergantung pada kemampuan lahan menahan air dan A adalah luas (satuan luas). Asumsi dalam rumus ini adalah *Intensitas hujan seragam diseluruh daerah dan Frekuensi banjir adalah sama seperti curah hujan.* Asumsi-asumsi ini tidak tepat untuk luasan yang luas, sehingga penggunaan rumus rasional pada luasan terbatas, seperti perencanaan drainase (bukan sungai).

$$Q = \frac{10^{-3} x 10^{6}}{3600} C_s CiA = \frac{C_s CiA}{3,6} \qquad \text{.......................... (3.2)}$$

dimana A dalam (km^2)

Rumus Rasional ditulis: $Qp = C_S C \frac{i\,A}{3,6}$ .... (3.3)

dimana:

Q = debit limpasan (m^3/dt)

C = koefisien pengaliran tergantung jenis permukaan

C_s = koefisien koreksi Rasional modifikasi

i = intensitas hujan rata-rata (mm/jam) untuk hujan deras yang durasinya sama dengan waktu konsentrasi (Tc), untuk mendapatkan i tersedia kurve IDF terlampir.

A = Luas daerah yang dikeringkan (km²)

Bila A dalam satuan (ha), maka rumus diatas menjadi:

$$Q = \frac{C_s CiA}{3,6.100} = 0,00278\, C_s CiA \qquad \text{............................ (3.4)}$$

dimana A dalam (ha)

Berdasarkan hasil modifikasi Rumus Rasional oleh beberapa ahli menetapkan; nilai koefisien pengaliran (C) berdasarkan pengaruh luasan daerah kedap air :

$$C = 0,9\frac{Ak}{A} + (1 + \frac{Ak}{A})Cp \qquad \text{............................ (3.5)}$$

dimana:

C = koefisien pengaliran

Ak = Luas daerah yang kedap (km, ha)

A = Luas total daerah yang akan di keringkan (km, ha)

Cp = koefisien pengaliran daerah yang tidak kedap

Tabel 3.1. Koefisien Limpasan (Manonobe's)

No	Jenis Permukaan Lahan	Koef. Limpasan C
1	Daerah pegunungan curam	0,75 – 0,90
2	Daerah pegunungan tersier	0,70 – 0,80
3	Tanah bergelombang dan hutan	0,50 – 0,75
4	Tanah dataran di tanami	0,45 – 0,60
5	Persawahan diairi	0,70 – 0,80
6	Sungai di daerah pegunungan	0,75 – 0,85
7	Sungai kecil di dataran	0,45 – 0,75
8	Sungai besar yang lebih dari 50% daerah	
	pengaliranya merupakan dataran	0,50 – 0,75

3.2. Formula IZZARD'S

Menetapkan time of equalibrium te dalam minutes untuk luasan kecil permukaan tanah tanpa saluran:

$$te = \frac{41\, bLo^{1/3}}{i^{2/3}} \qquad b = \frac{0{,}0007\ i + Cr}{So^{1/3}} \qquad \text{.........................(3.6)}$$

Lo = panjang aliran (feet), ***i*** = Intensitas hujan inches per hour dan So = slope permukaan.

Tabel 3.2. KOEFISIEN LIMPASAN DAN NILAI BANDING KEDAP AIR

Tata guna lahan	Karateristik	koef. Limpasan gabungan (C)	% Luas kedap	koef. Limpasan Lhn tdk kedap
1. Pusat bisnis dan pasar		0.90	100	0.90
2. Industri	penuh	0.91	80	0.95
3. Perumahan kepadatan	20 rmh/ha	0.61	30	0.48
tinggi	30 rmh/ha	0.69	40	0.55
	40 rmh/ha	0.80	60	0.65
	60 rmh/ha	0.86	75	0.75
4. Perumahan kepadatan	10 rmh/ha	0.50	20	0.40
rendah				
5. Taman	daerah datar	0.30	-	0.30
6. Lahan parkir	datar	0.86	95	-

Bila Lo dalam (m), intensitas hujan dalam (mm/jam), maka rumus te dalam menit, menjadi:

$$te = \frac{526\,bLo^{1/3}}{i^{2/3}} \qquad b = \frac{2{,}8x10^{-5}\,i + Cr}{So^{1/3}} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \ (3.7)$$

Rumus ini dapat digunakan untuk mendapatkan waktu konsentrasi tc luasan-luasan yang kecil pada permukaan tanah seperti; perumahan, jalan raya, pusat perdagangan, industri dll. Debit aliran dihitung dengan rumus rasional yang dikemukakan diatas.

Tabel 3.3. Retarding koefisien (Cr)

No	Permukaan	Nilai cr
1	Smooth asphalt surface	0,007
2	Concrete pavement	0,012
3	Tar and gravel pavement	0,017
4	Closly clipped sod	0,046
5	Dense bluegrass turf	0,060

Bila ditetapkan Intenstas hujan 4,5 inchi/jam, pada lapangan parkir sebuah bandara ukuran 150 ft x 160 ft, panjang lintasan air Lo = 34 ft, So = 0,001, permukaan asphal Cr = 0,007. Maka waktu konsentrasi aliran:

$$b = \frac{0,0007\ x 4,5 + 0,007}{0,001^{1/3}} = 0,102 \quad dan \quad te = \frac{41 x 0,102\ x 34^{1/3}}{4,5^{2/3}} = 4,98\ mnt$$

Waktu konsentrasi te = tc = 4,98 mnt dan Cs = 1, C = 0,90. Maka debit limpasan :

Tabel 3.4. Koefisien limpasan umum

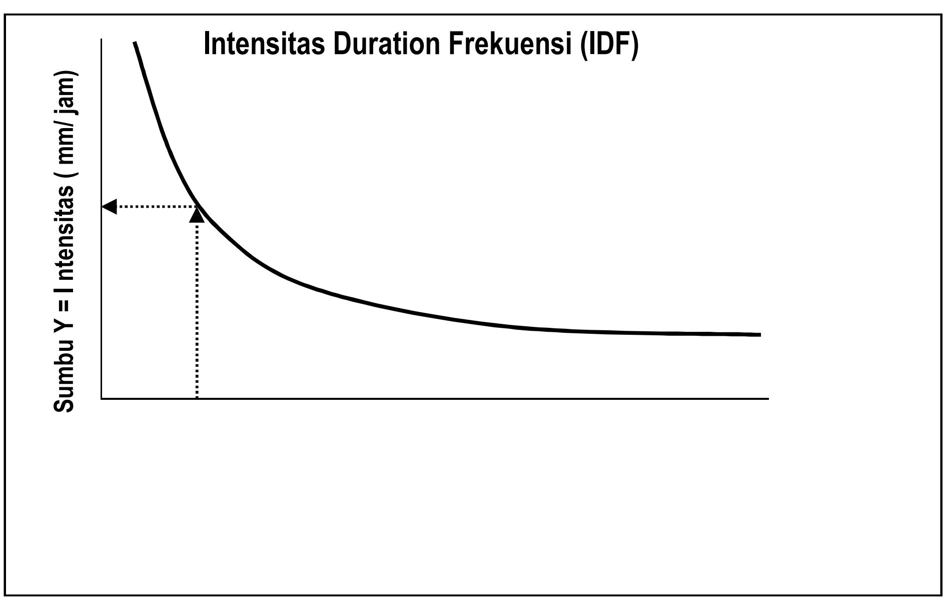
(Sumber buku Hidrolika, Ir. Iman Subarkah, 1978)

No	Jenis Lahan	% kemiringan	Loam berpasir	Lempung Silt loam	Lempung padat
1	Hutan	0-5	0,10	0,30	0,40
		5-10	0,25	0,35	0,50
		10-30	0,30	0,50	0,60
2	Padang rumput atau semak-semak	0-5	0,10	0,30	0,40
		5-10	0,15	0,35	0,55
		10-30	0,20	0,40	0,60
3	Tanah Pertanian	0 – 5	0,30	0,50	0,60
		5 – 10	0,40	0,60	0,70
		10 – 30	0,50	0,70	0,80

Cara mendapatkan nilai i_t (intensitas hujan)

a. Tentukan besar Tc pada titik tinjauan berdasarkan waktu aliran air dipermukaan dan saluran.

b. Dengan nilai Tc tersebut plotkan pada IDF di sumbu x, baca pada sumbu y besarnya intensitas hujan (i_t) yang sesuai



Curve Intensitas priode ulang tertentu

$$i_t = \frac{R_{24}}{24}\left(\frac{24}{Tc}\right)^{2/3}$$

Sumbu X = Tc (jam)

IDF dapat digunakan rumus dari Manonobe's

$$i_t = \frac{R_{24}}{24}\left(\frac{24}{Tc}\right)^{2/3} \qquad \text{..} \; (3.8)$$

dimana:

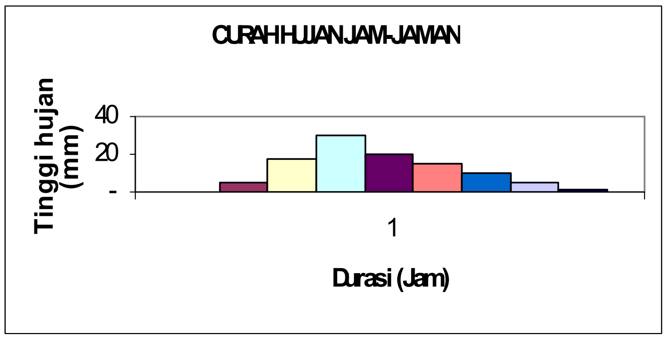
i_t = Intensitas hujan pada Tc tertentu (mm/jam)

R_{24} = Curah hujan sehari 24 jam (mm/hari)

T_c = waktu konsentrasi (jam)

Hujan untuk unit hidrograf dipilih curah hujan yang menimbulkan banjir misalnya dari pengamatan dilapangan ditemukan curah hujan efektif (setelah dukurangi dengan infiltrasi) 6, 12 dan 24 jam, menurut PSA-005 disusun berbentuk bell (genta). Hujan tersebut dialih ragamkan atau disuporposisi ke bentuk unit hidrograf, sehingga didapat hidrograf banjir rencana yang ditimbulkan hujan jam-jaman yang dipilih tersebut.

Jam	P(mm)
0	-
1	5.00
2	18.00
3	30.00
4	20.00
5	15.00
6	10.00
7	5.00



Koefisien koreksi tampungan (Cs) adalah perbandingan debit maksimum yang terjadi (Qmaks) dengan besarnya Intensitas hujan yang jatuh (i) pada luasan (A) dengan koefisien pengaliran (C).

Volume hujan yang jatuh:

$$V_1 = i.Tc.C.A \qquad (1)$$

Sedangkan dari hidrograf yang terjadi akibat hujan i selama Tc tersebut dapat ditentukan volume air yang terkumpul.

Volume air dari hidrograf:

$$V_2 = \frac{Qmaks}{2}(2Tc + Tt) \qquad (2)$$

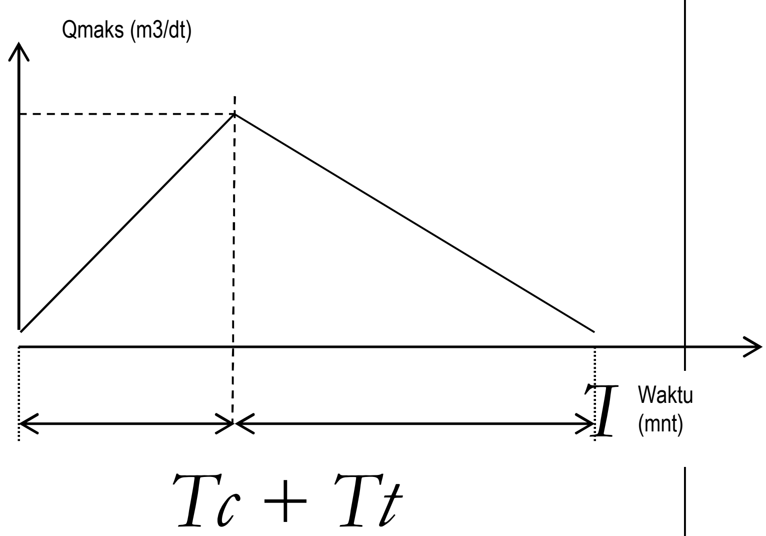
Pers (1) = (2), $V_1 = V_2$, maka:

$$Qmaks = i.\left[\frac{2Tc}{2Tc + Tt}\right].C.A$$

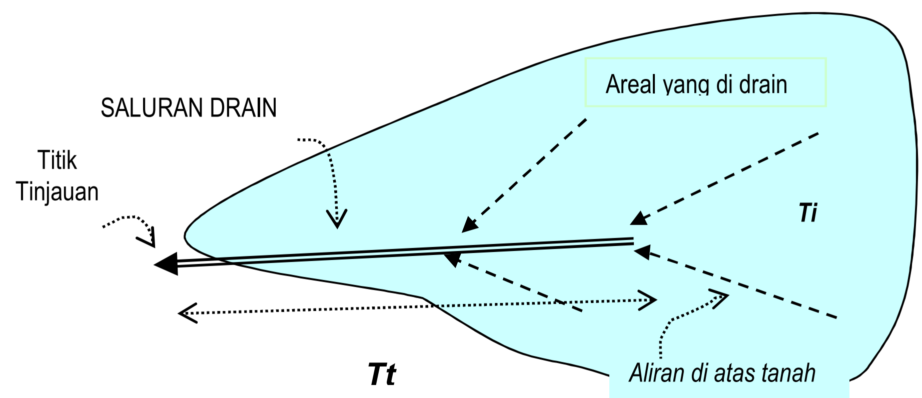
$$atau \quad \frac{2Tc}{2Tc + Tt} = \frac{Qmaks}{i.C.A}$$

Tc = Waktu konsentrasi = Ti + Tc
Ti = Waktu air mengalir dipermukaan tanah
Tt = Waktu air mengalir di saluran

Koefisien koreksi tampungan; $Cs = \frac{2Tc}{2Tc + Tt}$ .... (3.9)



HIDROGRAF METODE RASIONAL BERBENTUK SEGITIGA


Areal yang di drain
SALURAN DRAIN
Titik
Tinjauan
Ti
Tt
Aliran di atas tanah

$$Tc = Ti + Tt$$

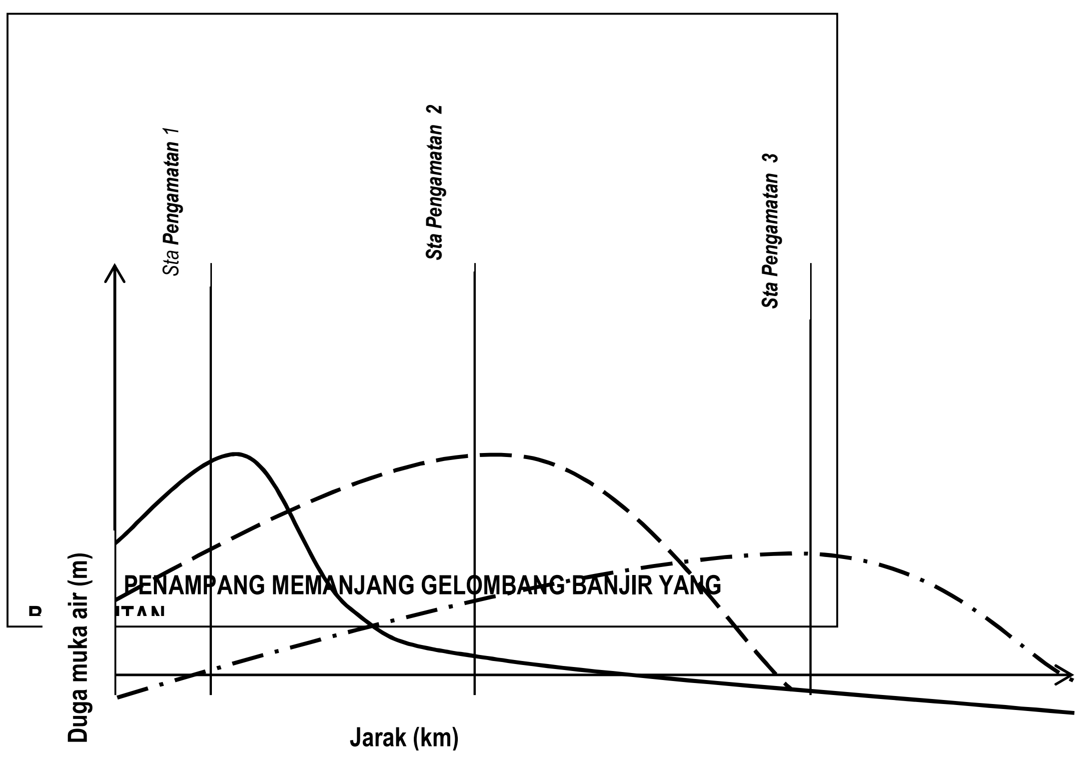
4 PENELUSURAN BANJIR (FLOOD ROUTING)

Rumus rasional adalah untuk luasan daerah pengaliran yang kecil, karena pengaruh tambahan volume air dalam perjalanan gelombang banjir yang pendek relatif kecil, lain halnya pada areal yang luas dan perjalanan gelombang banjir yang panjang.

Hidrograf adalah catatan tentang tinggi gelombang yang bergerak melalui suatu stasiun pengamatan. Bila gelombang tersebut bergerak arah kehilir, bentuknya akan berubah karena adanya tambahan volume aliran dari anak-anak sungai dan juga karena perpedaan kecepatan aliran sepanjang alur gelombang (lihat ilustrasi). Tanpa ada tambahan aliran, maka perubahan bentuk terjadi pemipihan atau penurunan tinggi puncak dan pemanjangan waktu gelombang banjir.

Maka dengan penjelasan diatas penggunaan rumus rasional dapat dijelaskan secara ilmiah tidak tepat bila digunakan untuk DAS lebih luas, karena hasil perhitungan debit banjir akan sangat besar.

Dengan demikian harus digunakan Metode Hidrograf Satuan Sintetik (HSS) untuk sungai-sungai yang tidak tersedia data pengamatan aliran yang cukup.



4.1. Waktu Konsentrasi (Tc)

Sesuai dengan penjelasan paragraf diatas penggunaan rumus rasional terbatas pada luasan yang kecil. Waktu konsentrasi untuk lahan yang alur-alur alirannya tidak kentara, seperti di DAS Lempuing disarankan menggunakan formula *IZZARD'S (Water Resources Engineering, Linsley & Franzini 1979), rumus ini berkaitan dengan intensitas hujan yang terjadi*.

Untuk saluran dan sungai digunakan formula untuk mendapatkan waktu konsentrasi formula dari *Kirpick*, dimana pengaruh panjang sungai dan kelandaian sungai berpengaruh penting, *Bransby-Williams* menambah pengaruh luasan DAS, penggunaanya harus dilakukan dengan pertimbangan cermat.

(1) *KIRPICH (1940)*

$$T_C = 0{,}01947\, L^{0,77}\, S^{-0,385} \quad \text{................} \quad (4.1)$$

Kirpich membuat rumus menggunakan pengaruh panjang sungai dan kelandaian.

Tc = waktu konsentrasi (menit)

L = panjang maksimum lintasan air (m)

S = kemiringan slope DAS = (ΔH / L)

(2) *BRANSBY-WILLIAMS*

$$Tc = \frac{0,604L}{A^{0,1}S^{0,2}} \qquad \text{..(4.2)}$$

L = panjang aliran (km), A = Luas catchment area (ha), S = Slope dan Tc dalam jam.

4.2. Analisa Kejadian Banjir

Analisa banjir menjadi penting guna menentukan konsep penanganan yang akan digunakan. Analisa penanganan banjir untuk debit banjir tertentu secara teknis ada dua kemungkinan:

1. Genangan banjir pada lahan, sungai tidak meluap
2. Genangan banjir pada lahan, sungai meluap

- *Genangan Banjir Pada Lahan, Sungai Tidak Meluap*

Untuk kasus ini bila hasil perhitungan debit banjir secara teoritis dialirkan pada palung sungai tidak menimbulkan peluapan, maka bila genangan masih terjadi, yang perlu diperbaiki adalah drainase dalam lahan.

- *Genangan Banjir Pada Lahan, Sungai Meluap*

Untuk kasus ini palung sungai tidak mampu menampung aliran, aliran dari lahan tertahan, sehingga menimbulkan genangan banjir. Maka usaha yang diperlukan adalah memperbaiki alur, menambah drain, menahan air sementara di lahan dan membangun tanggul. Karena semua air dari lahan tidak dapat dipaksa masuk kesungai pada waktu yang bersamaan, akan menimbulkan masalah pada bagian hilirnya. Maka perbuatan yang yang dilaksanakan parsial guna melancarkan aliran seperti sodetan, memperbesar tampang basah sungai tidak menyelesaikan masalah.

Untuk Sungai Lempuing adalah kasus yang kedua diatas maka untuk hal tersebut urutan kegiatan yang diperlukan adalah:

1. Pemeriksaan kapasitas tampang basah sungai berdasarkan debit banjir rencana tertentu.
2. Memeriksa ketinggian lahan-lahan yang pernah tergenang banjir dan kemampuan drainase lahan.
3. Merencanakan penanganan banjir.

4.3. Debit Banjir Sungai Lempuing

Data yang tersedia curah hujan harian maksimum dari 5 stasiun pengukur hujan dari tahun 1985 – 2001 yaitu stasiun Tugu Mulyo, Indralaya, Kayu agung, Dwisri dan Belitang. Perhitungan debit banjir menggunakan cara pengalih ragamkan curah hujan menjadi limpasan permukaan, melalui Metode Hidrograf Satuan Sintetik (HSS) cara ini merupakan cara empiris yang harus ditempuh karena tidak ada data pengukuran langsung yang cukup.

Kelemahan dari cara ini adalah hampir semua asumsi didasarkan pada pengalaman empiris di tempat lain. Cara kalibrasi dilakukan dengan cara membandingkan hasil perhitungan dengan pengukuran debit langsung, tentang kepantasan hasil perhitungan.

4.4. Curah Hujan Efektif

Pola curah hujan sangat berpengaruh kepada hidrograf yang dihasilkan, maka untuk memilih pola hujan yang menimbulkan banjir harus cermat. Curah hujan yang digunakan untuk analisis limpasan adalah curah hujan efektif atau disebut juga hujan mangkus yaitu curah hujan yang telah dikoreksi dengan faktor reduksi dan dikurangi dengan kehilangan seperti infiltrasi, intersepsi, penguapan dan tampungan cekungan.

Prof. Sri Harto (1989) & Barnes (1959), menganjurkan curah hujan yang digunakan untuk analisa harus dikurangi jumlah kehilangan air seperti; intersepsi, infiltrasi, penguapan dan *tampungan cekungan*. Disamping semua itu yang terpenting adalah Infiltrasi, perkiraan kehilangan air dalam suatu kasus *sangat sulit*, karena sangat dipengaruhi oleh tingkat kebasahan DAS (catchment wetness) sebelum tejadi hujan.

Maka berdasarkan penelitian banjir selama 10 tahun terakhir ini untuk penyederhanaan ditetapkan Indek ø yang bernilai tetap selama hujan berlangsung, formula ini digunakan pada areal yang dibatasi sampai 3.000 km^2. Karena semakin luas daerah jatuh hujan semakin besar air yang dapat diserap oleh lahan.

$$\phi = 10{,}4903 - 3{,}859 * 10^{-6} A + 1{,}6985 * 10^{-13} (A / SN)^4 \quad \text{..............(4.3)}$$

ø = indek infiltrasi (mm/jam)

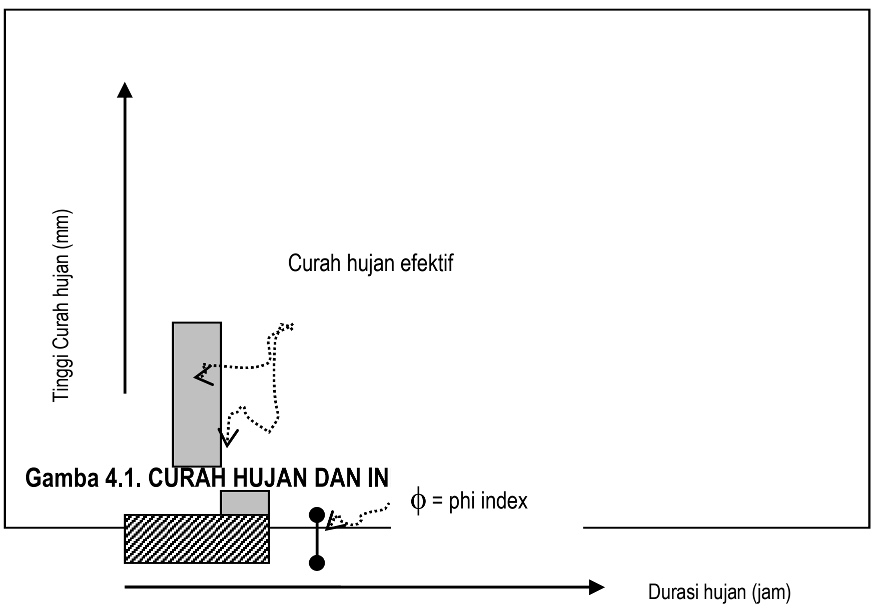
A = Luas daerah aliran sungai (km²)

SN = frekuesi sumber, tidak berdimensi yaitu jumlah banyak

Sungai tingkat 1dibandingkan jumlah seluruh segmen

sungai.

Cara mendapatkan curah hujan efektif dengan phi indeks adalah hasil curah hujan harian maksimum yang dari analisa frekuensi distribusi untuk return period tertentu dikurangi dengan besarnya phi indeks.



Gamba 4.1. CURAH HUJAN DAN IN

Hujan jam pertama belum menimbulkan limpasan permukaan, karena perlu mengisi kebasahan lahan. Terjadinya puncak banjir setelah terjadi penjenuhan lahan, air hujan dari tempat terjauh mengalir dipermukaan menuju tempat rendah dan mengalir di saluran atau sungai sampai ke tempat yang ditinjau, waktu pencapaian tempat tersebut disebut waktu konsentrasi (tc). Selama waktu tc infiltrasi dianggap tetap sebesar ø = indek infiltrasi (mm/jam). Hujan efektif adalah hujan yang telah dikurangi dengan indek infiltrasi dan koefisien koreksi lainya.

Sesuai dengan saran Prof. Sriharto, untuk DAS diluar Pulau Jawa diperlukan penyesuaian. Untuk itu kita harus tahu apa ciri-ciri DAS Pulau Jawa diantaranya:

1) Porsentase lahan terbuka lebih tinggi, hutanya berklasifikasi ringan. Artinya hujan yang jatuh cepat menuju sungai.
2) Debit sungai di Jawa perbedaan musim kering dan basah sangat besar, bahkan musim kering sungai tidak ada airnya.

4.5. Data Curah Hujan

Curah hujan yang tersedia dalam DAS Lempuing 5 stasuin diatas diperiksa keabsahanya dan selanjutnya di hitung return hujan 2, 5, 10, 25 , 50 dan 100 tahun dengan cara statistik menggunakan analisa frekuensi distribusi seperti; Pearson tipe III, Log Pearson tipe III, Log Normal dan Gumbel's.

Sumber hujan data Dinas PU Pengairan Sumatera Selatan, Volume 4 Data Book hasil The Study on Comprehensive Water Management Musi River Basin, 2003 dengan Grant JICA, Konsultan CTI Engineering Co. Ltd dan NIKKEN Consultant, Inc. Dimana data hasil pengamatan tersebut telah di uji keabsahanya melalui metode-metode hidrologi.

Hasil pengolahan secara statistik dengan beberapa metode Pearson tipe III, Log Pearson tipe III, Log Normal dan Gumbel's dengan hasil sebagai berikut:

Hasil analisa distribusi frekuensi 4 metode tersebut hampir sama berimpitan, sesuai dengan catatan penting tentang metode statistik untuk data maksimum yang lebih tepat hasilnya dari metode Gumbel's.

Tabel 4.1. Curah hujan harian mksimum DAS Lempuing

Lokasi	**Tugu Mulyo**		**Indralaya**		**Kayu Agung**		**Dewisri**		**Belitang**		h_{rt}
	mm	bln	mm	bln	mm	bln	mm	bln	mm	bln	mm
1985	128	Agust	127	Jul	77	Mei	106	Mart	124	Mart	112.40
1986	85	Agust	135	Apr	87	Mart	129	Sept	81	Sept	103.40
1987	141	Mart	55	Mart	103	Des	97	Mei	40	Mei	87.20
1988	127	Okt	121	Mart	81	Okt	125	Jan	50	Mart	100.80
1989	97	Des	180	Mei	92	Jan	85	Jul	117	Nop	114.20
1990	151	Mart	88	Jun	252	Agust	106	Jan	102	Mei	139.80
1991	104	Apr	86	Nop	115	Nop	83	Apr	147	Apr	107.00
1992	-		100	Jan	95	Jan	89	Okt	80	Mei	91.00
1993	139	Des	94	Apr	100	Apr	125	Feb	148	Jan	121.20
1994	50	Feb	120	Jan	96	Des	97	Jan	129	Des	98.40
1995	143	Mei	110	Apr	76	Des	79	Jan	147	Des	111.00
1996	92	Apr	119	Apr	90	Apr	-		-		100.33
1997	79	Jan	96	Des	97	Feb	91	Feb	69	Mei	86.40
1998	128	Jan	167	Sept	113	Okt	90	Des	97	Mart	119.00
1999	112	Jan	79	Mei	43	Des	133	Mei	162	Des	105.80
2000	120	Agust	91	Jan	61	Nop	75	Apr	83	Okt	86.00
2001	260	Sept	112	Apr	95	Mart	97	Mart	80	Des	128.80

Sumber data Study ComprehensiveWater Management Musi River Basin 2003

4.6. Faktor Reduksi dan Pola Distribusi Hujan

Panduan Perencanaan Bendungan Tipe Urugan, Ditjenair 1999 untuk Analisis Hidrologi dan PSA-007 menyarankan bagi daerah aliran yang tidak mempunyai cukup data pengamatan supaya memperhitungkan faktor reduksi dan bentuk distribusi hujan.

Faktor reduksi adalah faktor ketidak merataan hujan pada luasan DAS, untuk luas DAS 10 km^2 faktor reduksi = 1,0 dan semakin luas semakin kecil faktor reduksi 5.000 km^2 faktor reduksi 0,37.

Data hujan yang yang digunakan untuk perhitungan curah hujan periode ulang tertentu adalah data hujan harian maksimum, yang tidak diketahui polanya secara pasti (karena tidak ada data dari pengukur hujan otomatis). Maka berdasarkan pengalaman PSA-007 Ditjenair menyarankan, susunan distribusi hujan berbentuk bell. topi. Dimana hujan tertinggi ditempatkan ditengah, tertinggi kedua sebelah kiri, tertinggi ketiga sebelah kanan dan seterusnya.

Pemilihan durasi hujan dengan pola durasinya sangat berpengaruh pada hasil banjir rencana yang diperhitungkan. Curah hujan yang sama terdistribusi dengan durasi yang panjang akan menghasilkan puncak banjir lebih rendah dibandingkan dengan yang terdistribusi dengan durasi yang pendek.

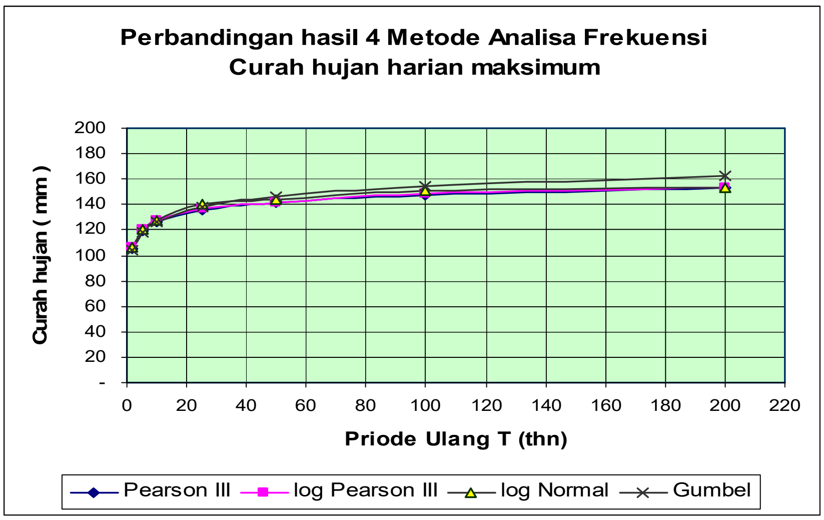
Oleh karena itu penetapan durasi hujan kritis perlu dilakukan optimasi beberapa besaran durasi hujan sehingga diperoleh durasi hujan yang kritis.Untuk bendungan-bendungan kecil disarankan durasi hujan 6 sampai 24 jam, misal 6, 9, 12, 15 dan seterusnya. Sedangkan untuk bendungan besar durasi hujan 1, 2, 3 hari bahkan dapat lebih tergantung besarnya DAS. Contoh hujan efektif R_{100}, CMB/PMF = Curah hujam Maksimum Boleh jadi.

Tabel 4.2. Perbandingan Hasil Perhitungan Curah Hujan

Analisa Frekuensi Curah hujan harian maksimum

Satuan (mm/hari)

Retr Priod	Analisa Frekuensi Distribusi				Keterangan
	Pearson III	log Pearson III	log Normal	Gumbel	
2	105.37	106.75	105.88	104.14	*dipilih yang*
5	118.88	120.00	119.90	117.54	*paling memenuhi*
10	126.68	127.97	127.97	126.42	*persyaratan*
25	135.58	136.99	140.31	137.63	
50	141.66	141.46	143.99	145.94	
100	147.34	148.44	150.69	154.20	
200	152.73	153.75	153.32	162.42	



Catatan Penting dalam menentukan frekuensi distribusi yang dipilih

1 Metode Gumbel lebih baik digunakan untuk analisis data maksimum seperti curah hujan, debit banjir

2 Metode Pearson III digunakan untuk analisis data maksimum yang bentuk kurvenya bell / lonceng / topi, mode terletak pada titik nol atau kemencenganyan Cs = 0

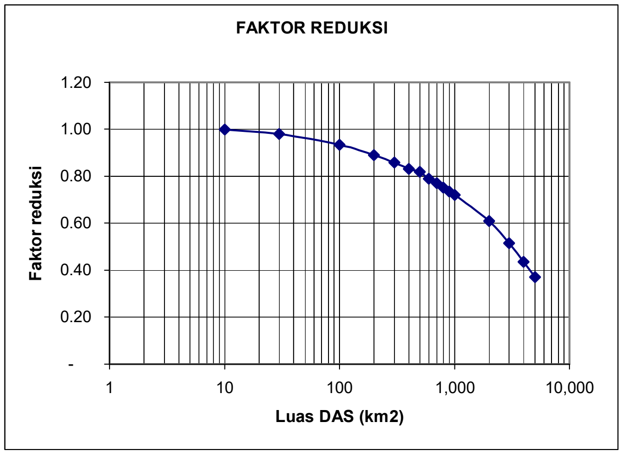
3 Metode Log Normal; Koefisien Kemencengan (Skewnees Koefisien) = C_s

Syarat distribusi Log Normal adalah ; $-0.1 < C_s < 0.1$

bila terlalu jauh mencengnya, tentu distribusi data ini tidak dapat dihitung dengan cara Log Normal. Artinya perkiraan yang dihasilkan agak jauh dari kebenaran

Tabel. 4.3. Faktor Reduksi

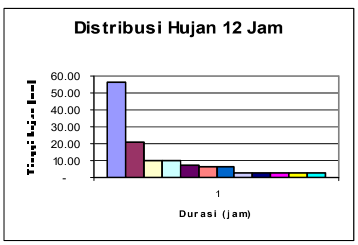
Luas DAS (km2)	Faktor Reduksi
10	1.00
30	0.98
100	0.94
200	0.89
300	0.86
400	0.83
500	0.82
600	0.79
700	0.77
800	0.75
900	0.74
1,000	0.72
2,000	0.61
3,000	0.52
4,000	0.44
5,000	0.37

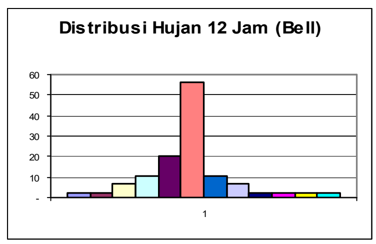


Tabel 4.4. Digunakan Distribusi hujan 12 jam

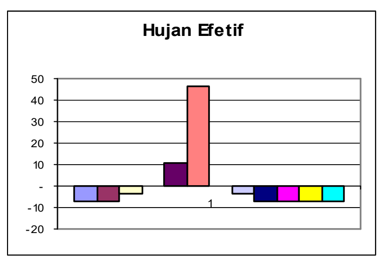
Koefisien Reduksi **0.83** **A = 409.00 km2**
Rt $_{100}$ (mm) **154.20 Curah Hujan Return Period 100 th**
Rn **127.99**

durasi hujan (jam)	intensitas porsentase cur. hujan CMB (%)	intensitas tinggi cur. hujan CMB (mm)	Distr. hujan jam2an CMB (mm)	Distr. hujan jam2an kritis (mm)	ϕ index (mm)	Distr. hujan jam2an efektif (mm)
a	b	c	d	e	f	g
1	44.00	56.31	56.31	2.56	9.91	(7.35)
2	60.00	76.79	20.48	2.56	9.91	(7.35)
3	68.00	87.03	10.24	6.40	9.91	(3.51)
4	76.00	97.27	10.24	10.24	9.91	0.33
5	82.00	104.95	7.68	20.48	9.91	10.57
6	87.00	111.35	6.40	56.31	9.91	46.40
7	90.00	115.19	6.40	10.24	9.91	0.33
8	92.00	117.75	2.56	6.40	9.91	(3.51)
9	94.00	120.31	2.56	2.56	9.91	(7.35)
10	96.00	122.87	2.56	2.56	9.91	(7.35)
11	98.00	125.43	2.56	2.56	9.91	(7.35)
12	100.00	127.99	2.56	2.56	9.91	(7.35)

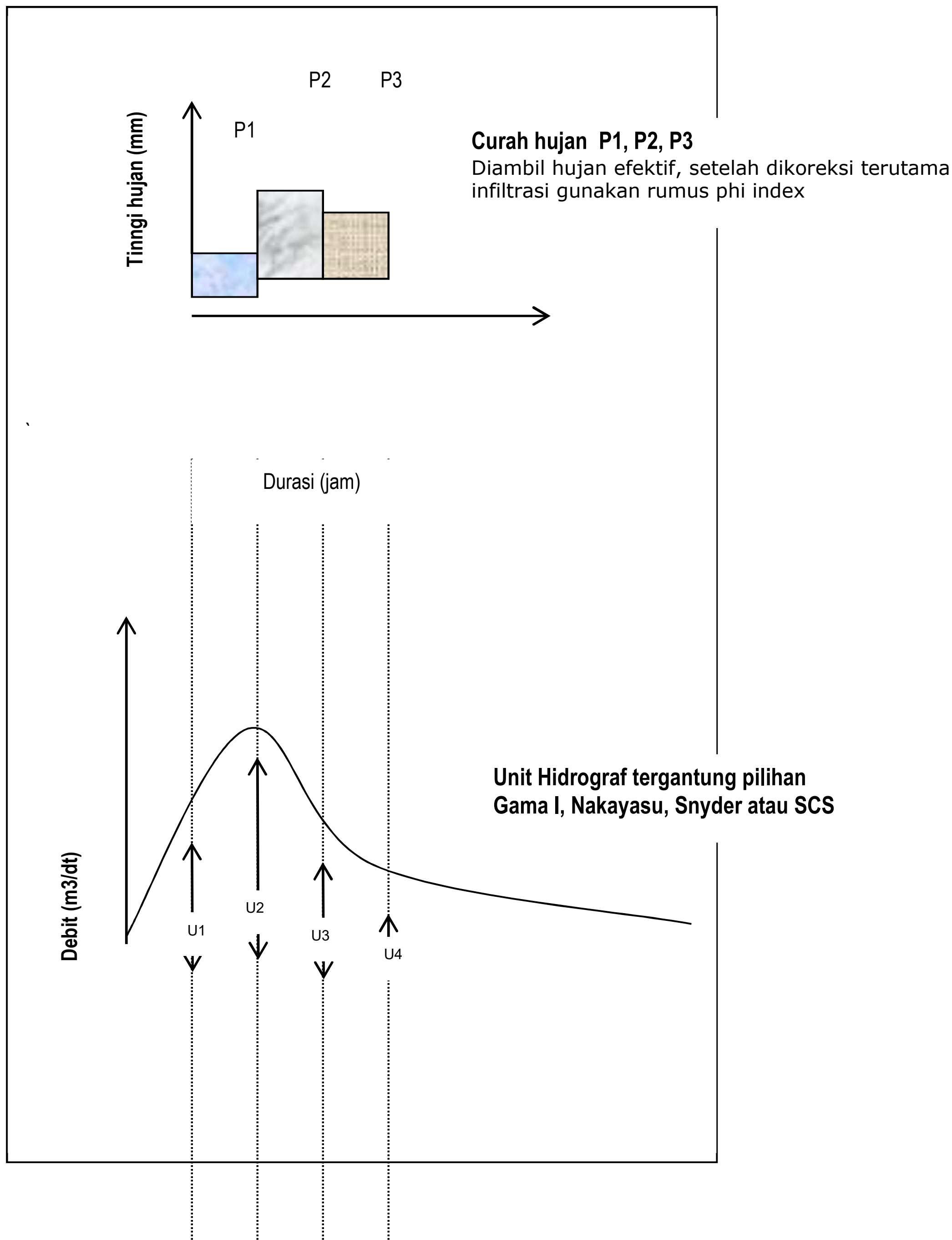




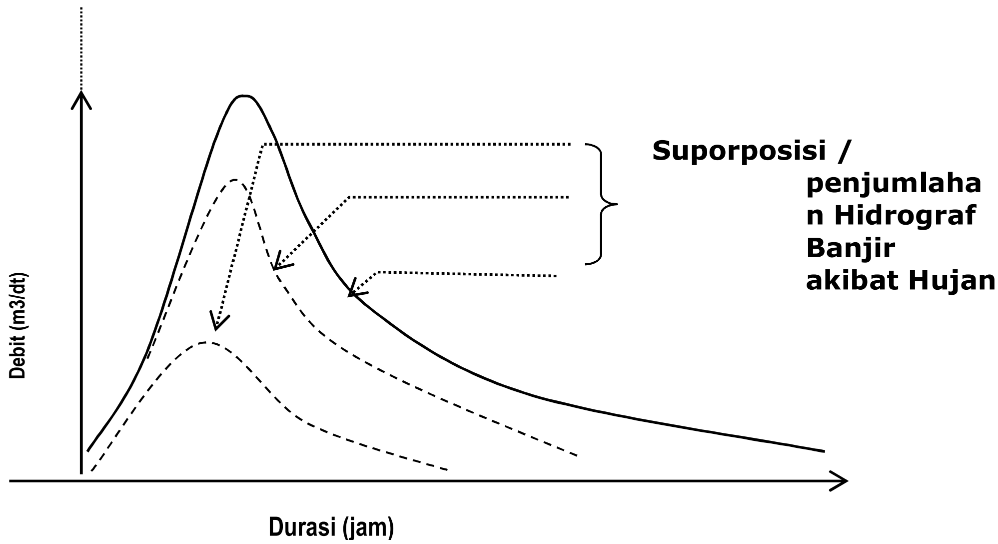
Keterangan:
angka dalam ( ) hujan tidak efektif
angka tidak berkurung hujan efektif



Untuk mendapatkan hidrograf banjir hujan efektif disuperposisikan ke Hidrograf Satuan Sintetik. Contoh hidrograf banjir akibat hujan efektif P1, P2 dan P3.



Durasi (jam)



Gambar 4.2. Cara pollynomial atau Collins

5 METODE HIDROGRAF SATUAN

Untuk sungai-sungai yang belum pernah diukur, cara mengatasi masalah dalam analisa hidrologi satu-satunya sampai saat ini adalah dengan *Analisa Hidrologi dengan memakai Hidrograf Satuan Sintetik.*

Akan dijelaskan beberapa teori Hidrograf Satuan Sintetik yang sering digunakan untuk menghitung debit banjir rencana.

(1) Metode Hidrograf Satuan Sintetik Gama I

(2) Metode Hidrograf Satuan Sintetik Nakayasu

(3) Metode Hidrograf Satuan Sintetik Snyder

(4) Metode Soil Concervatioan Services (SCS-USA)

Langkah-langkah perhitungan

1. Membuat Hidrograf Satuan Sintetik masing-masing sungai
2. Menghitung curah hujan efektif

3. Mensuporposisi curah hujan efektif ke HSS
4. Memilih metode yang sesuai dan cocok dengan bentuk hidrograf aliran hasil pengamatan pada sungai yang bersangkutan.

Tabel 5.1. Rumus metode HSS

	Debit Puncak & Waktu naik	Waktu konsentrasi
Gama I	Satuan kilometer, jam. $Qp = 0{,}1836\,A^{0,5886}\,JN^{0,2381}\,TR$ $TR = 0{,}43\,(\frac{L}{100\,SF})^{3} + 1{,}0665\,SIM$	Waktu konsentrasi / waktu naik diperhitungkan berdasarkan: panjang sungai L, faktor sumber SF, faktor simetri bentuk DAS disingkat SIM.

Nakayasu	Satuan kilometer, jam. $$Qp = \frac{CARo}{3{,}6(0{,}3Tp + T_{0,3})}$$ $Tp = (tg + 0{,}8.k.tg)$ $tg = 0{,}21\,L^{0.7} \ldots\ldots\ldots \ldots\ldots L < 15$ $tg = 0{,}4 + 0{,}058\,L \ldots\ldots\ldots L > 15$ $k = 0{,}5 \quad s/d \quad 1{,}0$	Waktu konsentrasi / waktu naik diperhitungkan berdasarkan: panjang sungai L, konstanta didasarkan pengalaman di Jepang. Dari pengalaman untuk sungai yang sama Tp Nakayasu lebih panjang dari Tp Gama I.
Snyder	Satuan mile, jam. $$Qp = 640\frac{Cp}{tp}A \qquad tp = Ct(L.Lc$$ $$te = \frac{tp}{5{,}5} \qquad Tb = 3 * 24 + \frac{tp}{8}$$	Waktu konsentrasi diperhitungkan berdasarkan: panjang sungai L, konstanta didasarkan pengalaman di AS. Dari pengalaman untuk sungai yang sama Tp Snyder lebih panjang dari Tp Nakayasu

SCS-USA	Satuan kilometer, jam. $Qp = 2{,}08\dfrac{A}{Tp} \quad (m^3/dt)$ $Tp = 0{,}60Tc + 0{,}5tr$ $Tc = 0{,}01947L^{0,77}S^{-0,385}$ $KIRPICH\ 1940 \quad Tc = (mnt),$	Waktu konsentrasi diperhitungkan berdasarkan: panjang sungai L, konstanta didasarkan pengalaman di AS. Dari pengalaman untuk sungai yang sama Tp SCS lebih panjang dari Tp Snyder.

5.1. Hidrograf Satuan Sintetik Gama I

Adalah hasil penelitian *Prof. Dr. Ir. Sriharto Br. Dipl. H,* terhadap 30 DAS di Pulau Jawa, rumus ini disarankan untuk luasan DAS tidak lebih dari 3.250 km2, dalam penelitian ditemukan 4 (empat) faktor yang dominan mempengaruhi proses terjadinya banjir yaitu:

a. Luas DAS (catchment area)
b. Panjang sungai (main stream length)
c. Landai sungai rata-rata (everage main strem slope)
d. Kerapatan jaringan kuras (drainage density)

Hal ini terungkap pula dalam penelitian lain hidrograf yang terukur di suatu stasion hidrometri berasal dari hujan yang jatuh dalam suatu DAS kemudian ditampung oleh sungai-sungai tingkat pertama (first order). Selanjutnya diteruskan ke sungai-sungai tingkat lebih besar sampai ke stasion hidrometri terukur hampir 80% berasal dari sungai-sungai tingkat satu.

Persandingan Gama I dengan hasil pengamatan lapangan dan hidrograf satuan sintetik metode lain, seperti Nakayasu, US–SCS, Common dan Snyder, ditemui penyimpangan cukup besar yang dapat mempengaruhi perhitungan debit yaitu pada;

(1) Waktu untuk pencapaian debit puncak (time lag = tp) Gama I lebih mendekati hasil lapangan yaitu lebih singkat, hidrograf metode lain lebih lama.

(2) Besarnya debit puncak (Qp) Gama I lebih tinggi sedikit dan hidrograf metode lain lebih rendah, metode ini lebih sesuai bila dibandingankan dengan hidrograf terukur dilapangan.

Kesimpulan hasil penelitian: Metode Gama I dan Nakayasu, lebih mendekati hasil pengamatan lapangan.

Debit puncak Qp (m3/dt)

$$Qp = 0{,}1836 \; A^{0{,}5886} \; JN^{0{,}2381} \; TR^{-0{,}4008} \quad \text{................. (5.1)}$$

A = Luas DAS (km2)

JN = Jumlah pertemuan sungai

TR = Waktu naik hidrograf (jam)

$$TR = 0{,}43\left(\frac{L}{100\,SF}\right)^{3} + 1{,}0665\,SIM + 1{,}2775 \quad \text{.................}$$

(5.2)

L = panjang sungai (km)

SF = factor sumber yaitu jumlah panjang semua sungai tingkat 1 dibandingkan dengan jumlah sungai semua tingkat.

SIM = factor simetri yaitu hasil kali antara faktor lebar (WF) dan

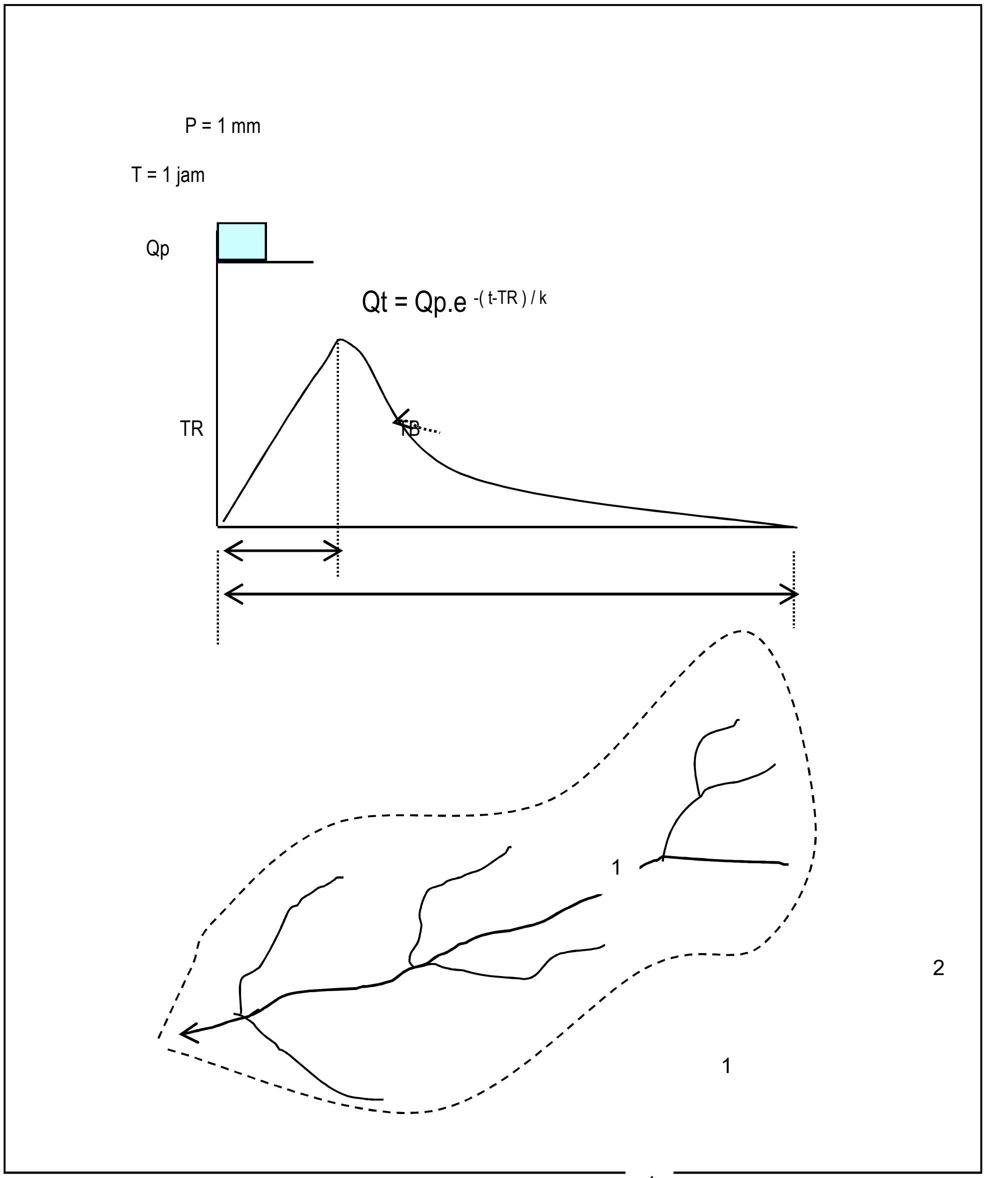
luas DAS sebelah hulu (RUA)

Setelah semua unsur-unsur dalam rumus dihitung dapat digamabarkan bentuk unit hidrograf Gama I:

1) Persamaan garis hidrograf menaik didekatkan dengan garis lurus
2) Persamaan hidrograf menurun bentuk persamaan exponensial

$Qt = Qp \, . e^{-t/K}$ dengan e = 2,718 .... (5.3)

Dalam persamaan Gama I Qp (debit puncak) telah memperhitungkan waktu konsentrasi atau waktu naik hidrograf (TR) berdasarkan panjang sungai induk (L) dan jumlah semua anak sungai yang berpengaruh (factor sumber = SF), bentuk simetri DAS (SIM).



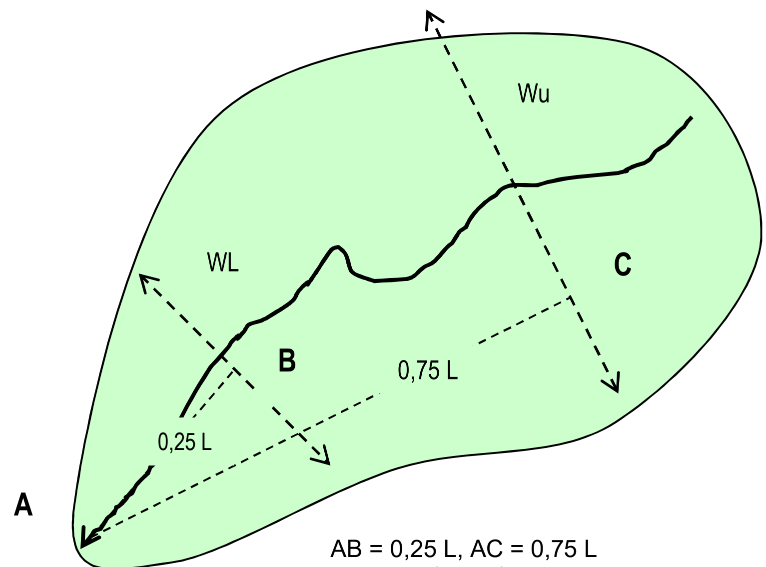
1

1

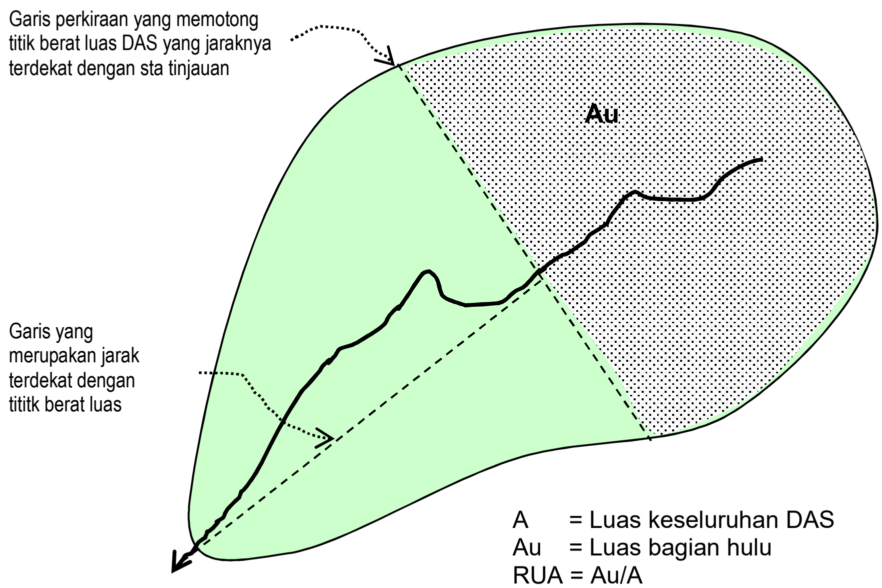
2

3

2



AB = 0,25 L, AC = 0,75 L
WF = Wu/WL, SIM = WF.RUA
L = panjang sungai



Wu = lebar DAS sejauh ¾ dari titik tinjauan

WL =lebar DAS sejauh ¼ dari titik tinjauan, dirumuskan:

$$WF = \frac{Wu}{WL} \qquad \text{....................... (5.4)}$$

WF = perbandingan lebar DAS sejauh 3/4 L dengan lebar DAS

Sejauh 1/4 Ldari titik peninjauan debit

RUA = luas DAS sebelah hulu ditentukan dengan cara

Memperkirakan letak tb (titik berat) luasan DAS dan buat garis pembatas yang merupakan garis tegak lurus terhadap garis jarak terdekat antara tb dan titik peninjauan debit.

Hidrograf satuan yang disajikan pada sisi naik merupakan garis lurus (TR) dan sisi turun merupakan garis dengan persamaan eksponensial (TD), dimana; waktu naik (TR) + turun (TD) = TB, sebagai berikut:

Persamaan lengkung debit HSS Gama I

$$Qt = Qp \ . e^{-t/K} \text{ dengan } e = 2{,}718 \qquad \text{................ (5.5)}$$

Qt = debit yang diukur pada jam ke t setelah debit puncak (m3/dt)

Qp= debit puncak (m3/dt)

K = koefisien tampungan (jam)

SF = Faktor sumber yaitu perbandingan antara jumlah panjang sungai tingkat 1 dengan jumlah panjang semua tingkat.

$$TB = 27{,}4132\, TR^{0{,}1457} . S^{-0{,}0986} . SN^{0{,}7344} . RUA^{0{,}2574}$$
$$K = 0{,}5617\, A^{0{,}1798} S^{-0{,}1446} SF^{-1{,}0897} D^{0{,}0452}$$

S = landai atau slope sungai rata-rata

D = kerapatan jaringan kuras yang merupakan panjang segmen seluruh tingkat sungai dibandingkan dengan luas DAS (km/km2)

Aliran dasar (base flow) yang berasal dari aliran air tanah. Besarnya aliran dasar yang dipengaruhi oleh luasnya DAS dan kerapatan jaringan kuras, rumus empiris ini mengandaikan besarnya aliran dasar tetap.

$$Qb = 0{,}4751\, A^{0{,}6444} D^{0{,}9430} \qquad \text{............................ (5.6)}$$

Untuk penggambaran hidrograf debit rencana harus ditambahkan aliran dasar Qb pada limpasan langsung.

Perhitungan kehilangan curah hujan di DAS dengan phi indeks:

$$\phi = 10,4903 - 3,859 * 10^{-6} A + 1,6985 * 10^{-13} (A / SN)^{4} \quad \text{............ (5.7)}$$

ø = indek infiltrasi (mm/jam)

A = Luas daerah aliran sungai (km^2)

SN = frekuesi sumber, tidak berdimensi yaitu jumlah banyak sungai tingkat 1dibandingkan jumlah seluruh segmen sungai.

Rumus ini sangat tergantung pada luas DAS, dalam penggunaan agar diperhatikan kebasahan DAS. DAS yang luas menghasilkan phi indeks yang tinggi.

5.2. Hidrograf Satuan Sintetik Nakayasu

Hidrograf Satuan Sintetik dasarnya adalah percobaan atau penelitian dilokasi dimana konstanta-konstanta rumus tersebut dibuat. Sedangkan bila diterapkan dilokasi lain perlu penyesuaian dengan karateristik lokasi, terutama yang berperan penting dalam rumus ini adalah:

(1) Panjang alur sungai (L), karena pengaruh L sangat besar pada perubahan time lag.

(2) Time lag yang merupakan bagian dalam fungsi bilangan pembagi pada rumus debit puncak (Qp).

Analisa hidrograf banjir metode Nakayasu menggunakan rumus rational dengan koefisien pengaliran dan konstanta ditetapkan berdasarkan empiris sebagai berikut:

Persamaan debit Nakayasu;

$$Qp = \frac{CARo}{3{,}6(0{,}3Tp + T_{0,3})} \qquad \text{......................} \ (5.8)$$

Rumus diatas telah memperhitungkan koef. pengaliran C, maka hujan efektif yang digunakan untuk penentuan debit banjir adalah hujan *sebelum dikurangi dengan phi index*.

dimana:

Q_P = debit puncak banjir (m3/dt)

C = koefisien pengaliran sangat tergantung kondisi vegetasi DAS

R_O = Curah hujan efektif (mm)

T_P = waktu dari permulaan hujan sampai puncak banjir (jam)

$T_{0,3}$ = waktu yang diperlukan oleh penurunan debit dari debit puncak sampai menjadi 30% debit puncak (jam)

Persamaan garis lengkung ke arah naik (rising limb) hidrograf satuan Nakayasu adalah:

$$Qa = Qp \left(\frac{t}{Tp}\right)^{2,4} \qquad \text{.................................} \quad (5.9)$$

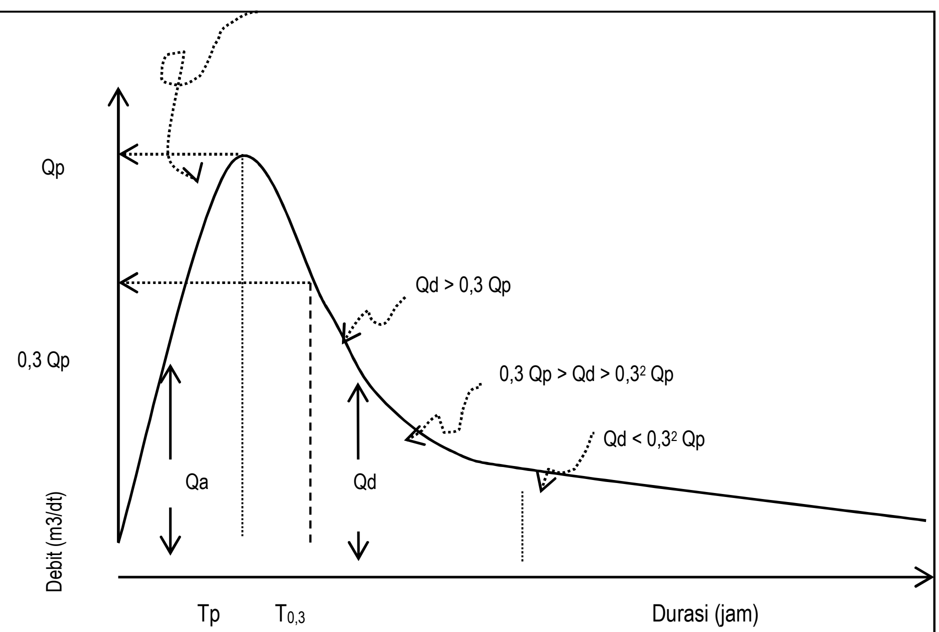
dimana:

Q_a = debit sebelum puncak dengan waktu t (m3/dt)

t = waktu dalam jam, diukur mulai titik 0 sampai t = Tp (waktu konsentrasi)

Bagian lengkung turun (decreasing limb); Q_d = debit bagian lengkung turun dibagi dalam 3 bagian persamaan, dimulai dari puncak debit mengikuti persamaan sebagai berikut :

$$Qa = Qp\left(\frac{1}{Tp}\right)^{2,4}$$



Gambar 5.4. Hidrograf Nakayasu

Bagian atas

- $Q_d > 0{,}3\ Q_P$: $Qd = Qp.0{,}3^{(\frac{t-Tp}{T_{0,3}})}$ ....(5.10)

Bagian tengah

- $0{,}3\ Q_P > Q_d > 0{,}3^2\ Q_P$: $Qd = Qp.0{,}3^{(\frac{t-Tp+0{,}5T_{0,3}}{1{,}5T_{0,3}})}$ ..... (5.11)

Bagian bawah

- $0{,}3^2\ Q_P > Q_d$: $Qd = Qp.0{,}3^{(\frac{t-Tp+1{,}5T_{0,3}}{2{,}0T_{0,3}})}$ ....(5.12)

pengambilan besarnya t dalam penyelesaian persamaan diukur dari titik nol 0 titik awal hidrograf.

dimana:

$T_P = tg + 0{,}8\ tr$

tr = satuan durasi hujan

tg tergantung panjang sungai (L)

- L < 15 km tg = 0,21. $L^{0,7}$
- L > 15 km tg = 0,4 + 0,058 L

dimana:

L = panjang alur sungai (km)

tg = waktu kosentrasi (jam)

tr = k tg (jam)

$k = 0{,}5 - 1{,}0$

$Tp = tg + 0{,}8\, k\, tg$

$Tp = tg\,(1 + 0{,}8\, k)$

$T_{0,3} = \alpha\, tg$ (jam)

Penjabaran rumus debit puncak Qp:

$$0{,}3Tp + T_{0,3} = 0{,}3(tg + 0{,}8.k.tg) + \alpha.tg$$
$$0{,}3Tp + T_{0,3} = (0{,}3 + 0{,}24.k + \alpha)tg$$

$$Qp = \frac{CARo}{3{,}6(0{,}3 + 0{,}24.k + \alpha)tg} \quad \text{satuan (m3/dt)} \qquad \text{......... (5.13)}$$

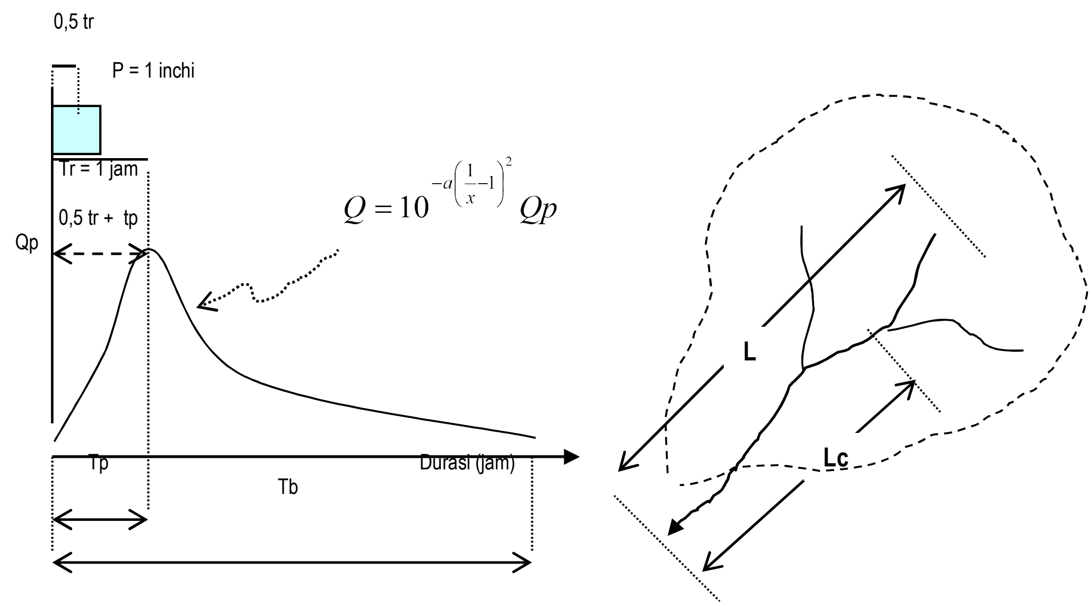
Besaran alpha (α) tergantung pada topografi daerah pengaliran dan kelandaian sungai dari percobaan di Jepang nilai sbb:

- Topografi dengan daerah pengaliran biasa α = 2, naik dan turun hidrograf seimbang.
- hidrograf yang naiknya lambat dan turunya cepat α = 1,5 digambarkan sebagai daerah pengaliran dengan topografi landai dan luas, bentuk segi empat atau agak bulat serta sungainya relatif lurus, agak terjal, tampang basah tidak tersedimentasi, sehingga aliran lancar.

- hidrograf yang naiknya cepat dan turunya lambat □ = 3 digambarkan sebagai sungai dengan daerah pengaliran agak terjal (daerah kaki pegunungan), tetapi alur sungainya lebih panjang didaerah yang relatif landai, sehingga aliran agak tertahan atau lambat.

5.3. Hidrograf Satuan Sintetik Snyder

Rumus ini dirancang dengan tinggi hujan P = 1 inchi dan waktu tr = 1 jam, jatuh pada luasan areal $mile^2$. Penelitian dilakukan oleh F.F.Snyder 1938 di banyak sungai Amerika Timur. Satuan yang digunakan untuk panjang sungai (L) mile, luas DAS dalam mile square ($mile^2$), time lag tp dalam jam sehingga debit puncak Qp dalam cubic feet per second.



Gambar 5.5. Hidrograf Snyder

Persamaan debit puncak Qp:

$$Qp = 640 \frac{Cp}{tp} A$$

$$tp = Ct(L.Lc)^{0,3}$$

$$te = \frac{tp}{5,5} \qquad \text{............................ (5.14)}$$

$$Tb = 3 * 24 + \frac{tp}{8}$$

L = panjang total sungai (mile)
Lc = panjang sungai dari titik berat DAS ke outlet (mile)

dimana :

Cp adalah koefisien yang dipengaruhi oleh waktu kelambatan (storage coefficient) nilai antara 0,56 s/d 0,69, *semakin besar nilai Cp waktu kelambatan semakin cepat, debit puncak semakin besar* dan Ct adalah koefisien yang dipengaruhi oleh kelandaian slope basin *semakin rendah nilainya semakin terjal* slope basin atau waktu konsentrasi pendek, puncak banjir cepat tercapai, Ct = antara 1,35 s/d 1,65.

Bila curah hujan P = 1 cm jatuh pada areal luasan km², tp (jam) dan debit puncak Qp dalam (m3/dt).

Maka : $Qp = \frac{1.000^2 x 10^{-2}}{3.600} \frac{Cp}{tp} A = 2,78 \frac{Cp}{tp} A$ .... (5.15)

Rumus diatas tidak menjelaskan tentang koef. pengaliran C, maka hujan efektif yang digunakan untuk penentuan debit banjir adalah hujan terukur *dikurangi dengan phi index*.

Pada penerapan akan terjadi beberapa kemungkinan besaran te:

- te < tr, dimana: tr = lama hujan efektif standar = 1 jam

  maka Tp = tp + 0,5 tr
- te > tr

  maka Tp = tp' + 0,5 tr, dimana: tp' = tp + 0,25 (tr – te)

Waktu dasar HSS : Tb = 72 + tp/8

Pengambarkan lengkung hidrograf mengikuti persamaan *Alexeyef* sebagai berikut:

$$Q = f(t) \qquad y = \frac{Q}{Qp} \quad atau \quad y = 10^{k} \qquad Q = 10^{-a\left(\frac{1}{x}-1\right)^{2}} Qp$$

$$x = \frac{t}{tp} \qquad k = -a(\frac{1}{x} - 1)^{2}$$

$$a = 1{,}32\lambda^{2} + 0{,}15\lambda + 0{,}045$$

$$\lambda = 3{,}6\frac{Qp.Tp}{h.A}$$

$$Qp = (m^{3}/dt), \quad h = (mm), \quad Tp = (jam), \quad A = (km^{2})$$

t = waktu yang dipilih untuk memudahkan membuat hidrograf (jam)

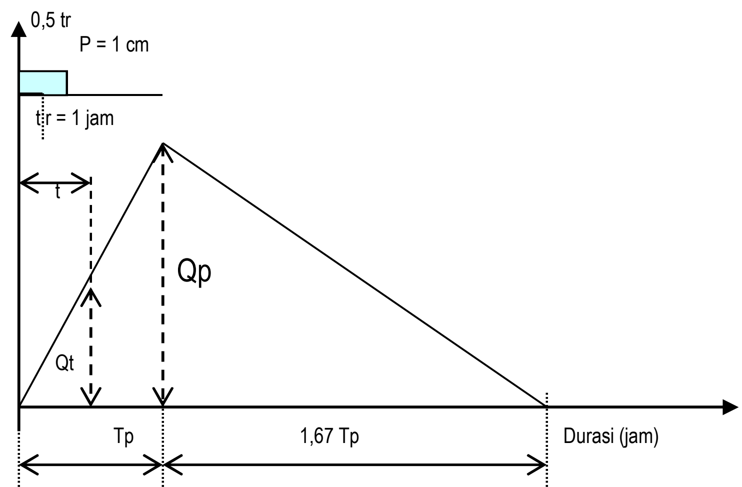
tp= waktu dari pertengahan terjadi hujan sampai puncak banjir (jam)

Urutan pembuatan unit hidrograf :

1. Tentukan L, Lc, A, tp, te, Qp, a dan Tb dengan h = 1 mm
2. tahap selanjutnya ambil besar t dimulai dari 0, 0,2, 1 dstnya, hitung nilai x = t/tp, λ dan a.
3. hitung harga k, selanjutnya hitung nilai setiap $Q_t = 10^k.Qp$, pada t tertentu dan selanjutnya unit hidrograf dapat digambarkan.

5.4. Hidrograf Satuan Sintetik SCS - USA

Hidrograf satuan Sintetik (SOIL CONCERVATION SERVICES) SCS–USA merupakan hidrograf satuan tidak berdimensi, di-expresikan dalam bentuk perbandingan debit q_t dengan debit puncak q_p dan waktu t dengan waktu naik (*time of rise*) t_p seperti tergambar. Nilai Qp dan Tp dapat diperkirakan dengan menggunakan penyederhanaan berbentuk setitiga. Dalam kajian terhadap banyak hidrograf satuan, waktu turun (*time of recession*) dapat diperkirakan sebesar 1,67 Tp dan basis hidrograf Tb = 2,67 Tp, Tp = tr/2 + 0,6 Tc dari Kirpich 1940. Untuk limpasan langsung (*direct runoff*) P = 1 inchi.



Gambar 5.6. Hidrograf SCS-USA

Persamaan Hidrograf Satuan Sintetik SCS-USA

HSS-SCS mengangap hidrograf berbentuk segitiga dengan ukuran seperti, volume hujan yang jatuh pada luasan A adalah:

$$Vol = \frac{Qp.Tp}{2} + \frac{Qp.1{,}67Tp}{2} = \frac{2{,}67}{2} Tp.Qp$$

$$Qp = \frac{2.Vol}{2{,}67Tp} = \frac{0{,}75.Vol}{Tp} \qquad \text{.............} \ (5.16)$$

A ($mile^2$), Tp (jam) dan Qp (cfs), 1 acre.inchi/jam = 1008 cfs

1 km = 0,621 mile, 1 mile = 5.280 ft, 1 $mile^2$ = 259 ha = 2,79 x 10^7 ft^2 = 2,59 km^2.

Disatu pihak, volume hujan yang jatuh sebesar 1 inchi, pada areal yang luasnya 1 mile, menghasilkan volume air sebesar :

$$Vol = \frac{5.280^2 x1/12}{3.600} A = 645{,}34.A$$

Maka persamaan diatas dapat diselesaikan

$$Qp = \frac{0{,}75(645{,}34)A}{Tp} = 484\frac{A}{Tp} \qquad (cfs) \ \ldots\ldots\ldots\ldots\ldots\ldots \quad (5.17)$$

dimana:

A dalam ($mile^2$), Tp dalam (jam), P = 1 inchi, Qp dalam (cfs).

Bila hujan yang jatuh 1 cm, pada areal yang luasnya 1 km^2, maka dihasilkan volume air sebesar:

$V = \dfrac{(1.000^2)x10^{-2}}{3600} A = 2{,}78A$, A dalam ($km^2$), Tp dalam (jam).

$$Qp = \frac{0{,}75(2{,}78)A}{Tp} = 2{,}08\frac{A}{Tp} \qquad (m^3/dt) \ \ldots\ldots\ldots\ldots\ldots\ldots \quad (5.18)$$

Maka formula debit puncak: $Qp = 2{,}08\dfrac{A}{Tp} \qquad (m^3/dt)$

$$Tp = 0{,}60Tc + 0{,}5tr$$

$$Tc = 0{,}01947L^{0,77}S^{-0,385}$$

$$KIRPICH\ 1940 \qquad Tc = (mnt), \qquad L = (m)$$

q naik mengikuti rumus $Qt = \frac{t}{Tp}Qp$

q turun mengikuti rumus $Qt = \frac{t}{1{,}67Tp}Qp$

Waktu dasar Tb = 2,67 Tp

tr = waktu hujan standar = 1 jam, dengan tinggi hujan P = 1 cm.

Rumus Qp diatas belum memperhitungkan koef. pengaliran C, maka hujan efektif yang digunakan untuk penentuan debit banjir adalah hujan *setelah dikurangi dengan phi index*.

5.5. Hasil Perhitungan Debit Banjir

Perhitungan debit dilakukan dengan 4 metode hidrograf satuan sintetik, gunanya untuk sebagai perbandingan dalam penetapan besarnya debit rencana. Dibawah ini hasil dari perhitungan debit untuk periode ulang 2, 5, 10, 25, 50 dan 100 tahun.

Berdasarkan Pedoman Pengendalian Banjir Volume 1,

Direktorat Jenderal Pengairan, Februari 1996.

Tabel 5.2. Kala Ulang Minimum yang Disarankan sebagai Banjir RencanaYang Berkenaan dengan Genangan Banjir

Sistem Aliran	• **Didasarkan pada tipe proyek Pengendalian Banjir** • **Didasarkan pada populasi penduduk**	**Fase awal**	**Fase Akhir**
Sungai	- Proyek Darurat - Proyek Baru - Pedesaan atau kota dengan P< 2 juta - Untuk perkotaan dengan P> 2 juta	5 10 25 25	10 25 50 100
Sistem Drainase Primer DAS < 500 ha	- Perdesaan - Perkotaan P < 0,5 juta - Perkotaan 0,5 juta < P < 2 juta - Perkotaan P > 2 juta	2 5 5 10	5 10 15 25
Sistem Drainase Sekunder DAS < 500 ha	- Perdesaan - Perkotaan P < 0,5 juta - Perkotaan 0,5 juta < P < 2 juta - Perkotaan P > 2 juta	1 2 2 5	2 5 5 10
Sistem Drainase Tertier DAS < 10 ha	- Perkotaan dan Perdesaan	1	2

Maka untuk kriteria Desain Pengendalian Banjir Sub DAS Lempuing digunakan fase awal sebagai proyek baru Periode Ulang 10 tahun untuk sungai dan 2 tahun untuk sistem drainase. Debit Rencana dipilih berdasarkan perhitungan Gama I untuk anak sungai dan Nakayasu untuk sungai Lempuing.

Tabel 5.3. Debit Banjir Rencana pada DAS Lempuing

No	Nama Sungai	Debit Banjir Q_{10} (m^3/dt)
1	Sungai Burnai	378,59
2	Sungai Macak	407,64
3	Sungai Belitang	420,86
4	Sungai Way Hitam	467,70
5	Sungai Lempuing SL.1	859.70
6	Sungai Lempuing SL.2	1076,59
7	Sungai Lempuing SL.3	1391,01

Sumber: hasil analisis penulis, 2010

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Selain mengajar juga aktif di berbagai organisasi profesi antara lain HATHI (Himpunan Ahli Teknik Hidraulik Indonesia), KNI-ICID, PII (Persatuan Insinyur Indonesai) dan LPJK (Lembaga Pengembang Jasa Konstruksi Indonesia) Provinsi Sumatera Selatan.

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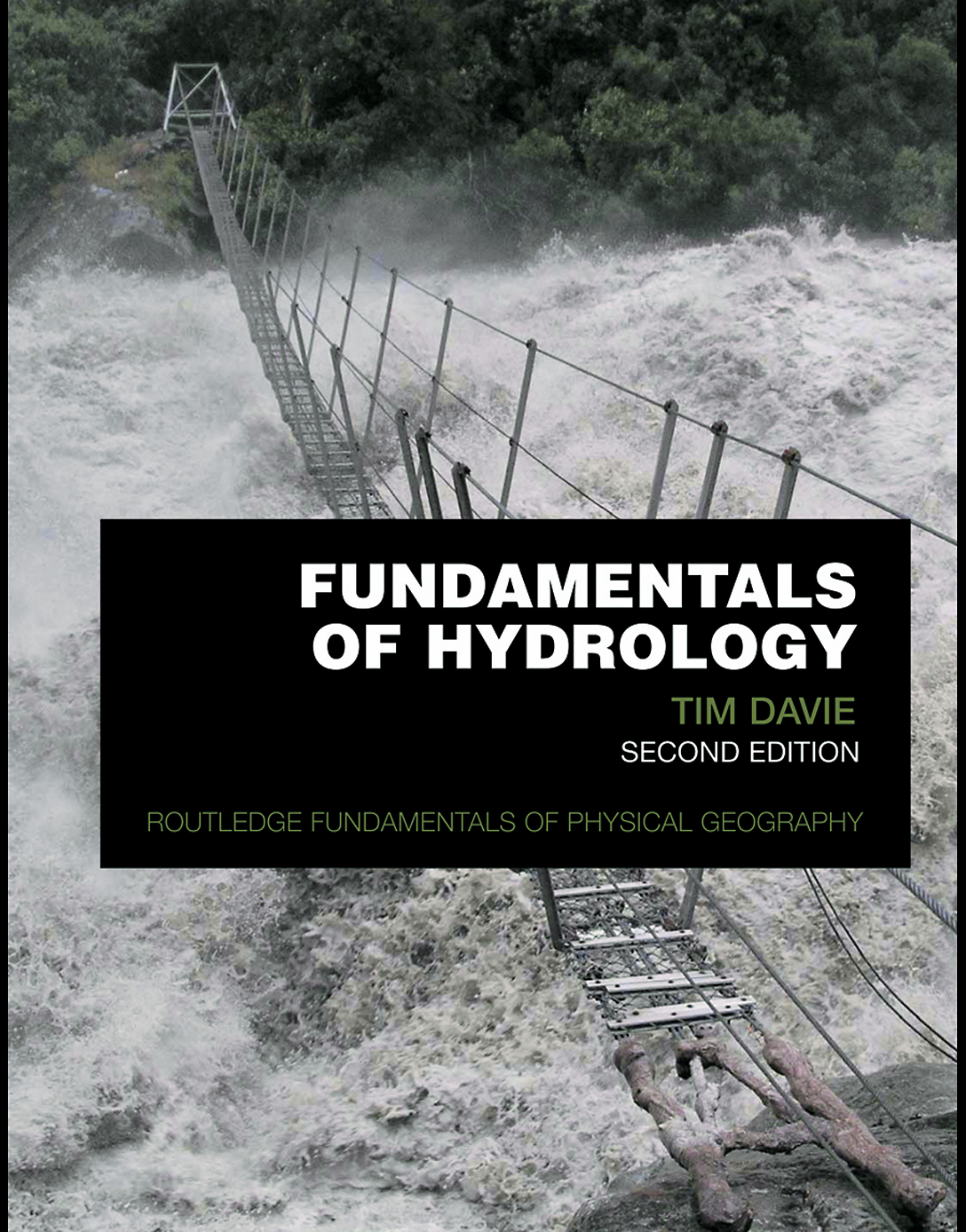
Lempuing (2005), Perkuatan tebing pengendalian banjir sungai Musi (2016)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Tim Davie is a research scientist working in the areas of land use change hydrology and Integrated Catchment Management in New Zealand. He is President of the New Zealand Hydrological Society and previously lectured in Environmental Science and Geography at Queen Mary College, University of London.

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To Christine, Katherine and Sarah Davie

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SERIES EDITOR'S PREFACE

We are presently living in a time of unparalleled change and when concern for the environment has never been greater. Global warming and climate change, possible rising sea levels, deforestation, desertification, and widespread soil erosion are just some of the issues of current concern. Although it is the role of human activity in such issues that is of most concern, this activity affects the operation of the natural processes that occur within the physical environment. Most of these processes and their effects are taught and researched within the academic discipline of physical geography. A knowledge and understanding of physical geography, and all it entails, is vitally important.

It is the aim of this *Fundamentals of Physical Geography Series* to provide, in five volumes, the fundamental nature of the physical processes that act on or just above the surface of the earth. The volumes in the series are *Climatology*, *Geomorphology*, *Biogeography*, *Hydrology* and *Soils*. The topics are treated in sufficient breadth and depth to provide the coverage expected in a *Fundamentals* series. Each volume leads into the topic by outlining the approach adopted. This is important because there may be several ways of approaching individual topics. Although each volume is complete in itself, there are many explicit and implicit references to the topics covered in the other volumes. Thus, the five volumes together provide a comprehensive insight into the totality that is Physical Geography.

The flexibility provided by separate volumes has been designed to meet the demand created by the variety of courses currently operating in higher education institutions. The advent of modular courses has meant that physical geography is now rarely taught in its entirety in an 'all-embracing' course, but is generally split into its main components. This is also the case with many Advanced Level syllabuses. Thus students and teachers are being frustrated increasingly by the lack of suitable books and are having to recommend texts of which only a small part might be relevant to their needs. Such texts also tend to lack the detail required. It is the aim of this series to provide individual volumes of sufficient breadth and depth to fulfil new demands. The volumes should also be of use to sixth form teachers where modular syllabuses are also becoming common.

Each volume has been written by higher education teachers with a wealth of experience in all aspects of the topics they cover and a proven ability in presenting information in a lively and interesting way. Each volume provides a comprehensive coverage of the subject matter using clear text divided into easily accessible sections and subsections. Tables, figures and photographs are used where appropriate as well as boxed case studies and summary notes. References to important previous studies and results are included but are used sparingly to avoid overloading the text. Suggestions for further reading are also provided. The main target readership is introductory level undergraduate students of physical geography or environmental science, but there will be much of interest to students from other disciplines and it is also hoped that sixth form teachers will be able to use the information that is provided in each volume.

John Gerrard

AUTHOR'S PREFACE
(First Edition)

It is the presence or absence of water that by and large determines how and where humans are able to live. This in itself makes water an important compound, but when you add in that the availability of water varies enormously in time and space, and that water is an odd substance in terms of its physical and chemical properties, it is possible to see that water is a truly extraordinary substance worthy of study at great length. To study **hydrology** is to try and understand the distribution and movement of fresh water around the globe. It is of fundamental importance to a rapidly growing world population that we understand the controls on availability of fresh water. To achieve this we need to know the fundamentals of hydrology as a science. From this position it is possible to move forward towards the management of water resources to benefit people in the many areas of the world where water availability is stressed.

There have been, and are, many excellent textbooks on hydrology. This book does not set out to eclipse all others, rather it is an attempt to fit into a niche that the author has found hard to fill in his teaching of hydrology in an undergraduate Physical Geography and Environmental Science setting. It aims to provide a solid foundation in the fundamental concepts that need to be understood by anybody taking the study of hydrology further. These fundamental concepts are: an understanding of process; an understanding of measurement and estimation techniques; how to interpret and analyse hydrological data; and some of the major issues of change confronting hydrology. One particular aspect that the author has found difficult to find within a single text has been the integration of water quantity and quality assessment; this is attempted here. The book is aimed at first- and second-year undergraduate students.

This book also aims to provide an up-to-date view on the fundamentals of hydrology, as instrumentation and analysis tools are changing rapidly with advancing technology. As an undergraduate studying physical geography during the 1980s, an older student once remarked to me on the wisdom of studying hydrology. There will be very little need for hydrologists soon, was his line of thought, as computers will be doing all the hydrological analysis necessary. In the intervening twenty years there has been a huge growth in the use of computers, but fortunately his prediction has turned out to be incorrect. There is a great need for hydrologists – to interpret the mass of computer-generated information, if nothing else. Hydrology has always been a fairly numerate discipline and this has not changed, but it is important that hydrologists understand the significance of the numbers and the fundamental processes underlying their generation.

There is an undoubted bias in this book towards the description of hydrology in humid, temperate regions. This is a reflection of two factors: that the author's main research has been in the UK and New Zealand, and that the majority of hydrological research has been carried out in humid and temperate environments. Neither of these is an adequate excuse to ignore arid regions or those dominated by snow and ice melt, and I have tried to incorporate some description of processes relevant to these environs. The book is an attempt to look at the fundamentals of hydrology irrespective of region or physical environment, but it is inevitable that some bias does creep in; I hope it is not to the detriment of the book overall.

There are many people whom I would like to thank for their input into this book. In common with many New Zealand hydrologists it was Dave Murray who sparked my initial interest in the subject and has provided many interesting discussions since. At the University of Bristol, Malcolm Anderson introduced and guided me in the application of modelling as an investigative technique. Since then numerous colleagues and hydrological acquaintances have contributed enormously in enhancing my understanding of hydrology. I thank them all. Keith Smith initially suggested I write this text, I think that I should thank him for that! The reviewers of my very rough draft provided some extremely constructive and useful criticism, which I have tried to take on board in the final version. I would particularly like to thank Dr Andrew Black from the University of Dundee who commented on the initial proposal and suggested the inclusion of the final chapter. Thanks to Ed Oliver who drew many of the diagrams. My wife Chris, and daughters Katherine and Sarah, deserve fulsome praise for putting up with me as I worried and fretted my way past many a deadline while writing this.

Tim Davie
London, December 2001

AUTHOR'S PREFACE
(Second Edition)

In the first edition of *Fundamentals of Hydrology* I started by pointing out the importance of hydrology as a science. I am sure all scientists could, and do, point out the same thing for their discipline. The reason I was first drawn to hydrology above other scientific disciplines was to understand the processes that lead to water flowing down a river. I wanted to know where the water flowing down a river had come from and how long it had taken to get there. I also have a social consciousness that wanted satisfaction in knowing that my learning was useful to people. As a University Lecturer from 1993–2001, in addition to research, I spent a lot of time sharing my passion for hydrological understanding through teaching. This culminated in my writing the first edition of *Fundamentals of Hydrology*, which was to fill a need I found in linking of water quantity and quality. Since the publication of the first edition I have been working as a scientist in a multi-disciplinary environment with a strong focus on applied research: science that directly benefits end-users. With this in mind, the second edition of *Fundamentals of Hydrology* has included extra sections on water resource management concepts and some of the linkages between ecology and hydrology. This edition has also benefited from the feedback provided by readers and reviewers. In response to this feedback the text has been rewritten to a slightly higher level and there are more illustrations and case studies. The chapter structure has been simplified with the text around rainfall interception (Chapter 4 in the first edition) being incorporated within the precipitation and evaporation chapters. I have also attempted to integrate the water quality and quantity aspects of hydrology to a greater degree through the addition of extra sections linking the physical processes with water quality.

The second edition also provides an updated version of hydrological science. Hydrological knowledge is increasing and there is a constant need to update any text book in light of recent discoveries. In the second edition of *Fundamentals of Hydrology* there are over fifty new references and each chapter has been reviewed in light of recent research findings.

In addition to a changed working environment, the new edition of the book has benefited from many informal discussions on hydrological matters that I have been able to have while at work. In particular I would like to thank Barry Fahey, Rick Jackson, Andrew Fenemor, Joseph Thomas and Mike Bonell for sharing their considerable insights with me. I am grateful to my employer, Landcare Research NZ Ltd, which has generously allowed me the time to finish this second edition through the provision of Capability Funding (from the New Zealand Foundation for Research Science and Technology). Those that were acknowledged in the first edition remain in my mind as important components of this book's evolution. In particular I think of Dave Murray who has died since the publication of the first edition. The staff at Routledge, and in particular Andrew Mould and Jennifer Page, have been extremely tolerant of my idiosyncrasies, I thank them for that. I remain particularly grateful to my wife Chris and children, Katherine and Sarah, who once again have put up with me as I work my way past deadlines but also are subjected to many impromptu hydrological lessons as we travel on holidays.

Tim Davie
Lincoln, New Zealand, August 2007

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HYDROLOGY AS A SCIENCE

Quite literally hydrology is 'the science or study of' ('logy' from Latin *logia*) 'water' ('hydro' from Greek *hudor*). However, contemporary hydrology does not study all the properties of water. Modern hydrology is concerned with the distribution of water on the surface of the earth and its movement over and beneath the surface, and through the atmosphere. This wide-ranging definition suggests that all water comes under the remit of a hydrologist, while in reality it is the study of fresh water that is of primary concern. The study of the saline water on earth is carried out in oceanography.

When studying the distribution and movement of water it is inevitable that the role of human interaction comes into play. Although human needs for water are not the only motivating force in a desire to understand hydrology, they are probably the strongest. This book attempts to integrate the physical processes of hydrology with an understanding of human interaction with fresh water. The human interaction can take the form of water quantity problems (e.g. over-extraction of **groundwater**) or water quality issues (e.g. disposal of pollutants).

> Water is among the most essential requisites that nature provides to sustain life for plants, animals and humans. The total quantity of fresh water on earth could satisfy all the needs of the human population if it were evenly distributed and accessible.
>
> (Stumm, 1986: p201)

Although written over twenty years ago, the views expressed by Stumm are still apt today. The real point of Stumm's statement is that water on earth is not evenly distributed and is not evenly accessible. It is the purpose of hydrology as a pure science to explore those disparities and try and explain them. It is the aim of hydrology as an applied science to take the knowledge of why any disparities exist and try to lessen the impact of them. There is much more to hydrology than just supplying water for human needs (e.g. studying floods as natural hazards; the investigation of lakes and rivers for ecological habitats), but analysis of this quotation gives good grounds for looking at different approaches to the study of hydrology.

The two main pathways to the study of hydrology come from engineering and geography, particularly the earth science side of geography. The earth science approach comes from the study of landforms (**geomorphology**) and is rooted in a history of explaining the processes that lead to water moving around the earth and to try to understand spatial links between the processes. The engineering approach tends to be a little more practically based and is looking towards finding solutions to problems posed by water moving (or not moving) around the earth. In reality there are huge areas of overlap between the two and it is often difficult to separate them, particularly when you enter into hydrological research. At an undergraduate level, however, the difference manifests itself through earth science hydrology being more descriptive and engineering hydrology being more numerate.

The approach taken in this book is more towards the earth science side, a reflection of the author's training and interests, but it is inevitable that there is considerable crossover. There are parts of the book that describe numerical techniques of fundamental importance to any practising hydrologist from whatever background, and it is hoped that the book can be used by all undergraduate students of hydrology.

Throughout the book there are highlighted case studies to illustrate different points made in the text. The case studies are drawn from research projects or different hydrological events around the world and are aimed at reinforcing the text elsewhere in the same chapter. Where appropriate, there are highlighted worked examples illustrating the use of a particular technique on a real data set.

IMPORTANCE OF WATER

Water is the most common substance on the surface of the earth, with the oceans covering over 70 per cent of the planet. Water is one of the few substances that can be found in all three states (i.e. gas, liquid and solid) within the earth's climatic range. The very presence of water in all three forms makes it possible for the earth to have a climate that is habitable for life forms: water acts as a *climate ameliorator* through the energy absorbed and released during transformation between the different phases. In addition to lessening climatic extremes the transformation of water between gas, liquid and solid phases is vital for the transfer of energy around the globe: moving energy from the equatorial regions towards the poles. The low viscosity of water makes it an extremely efficient transport agent, whether through international shipping or river and canal navigation. These characteristics can be described as the *physical properties* of water and they are critical for human survival on planet earth.

The *chemical properties* of water are equally important for our everyday existence. Water is one of the best solvents naturally occurring on the planet. This makes water vital for cleanliness: we use it for washing but also for the disposal of pollutants. The solvent properties of water allow the uptake of vital nutrients from the soil and into plants; this then allows the transfer of the nutrients within a plant's structure. The ability of water to dissolve gases such as oxygen allows life to be sustained within bodies of water such as rivers, lakes and oceans.

The capability of water to support life goes beyond bodies of water; the human body is composed of around 60 per cent water. The majority of this water is within cells, but there is a significant proportion (around 34 per cent) that moves around the body carrying dissolved chemicals which are vital for sustaining our lives (Ross and Wilson, 1981). Our bodies can store up energy reserves that allow us to survive without food for weeks but not more than days without water.

There are many other ways that water affects our very being. In places such as Norway, parts of the USA and New Zealand energy generation for domestic and industrial consumption is through hydro-electric schemes, harnessing the combination of water and gravity in a (by and large) sustainable manner. Water plays a large part in the spiritual lives of millions of people. In Christianity baptism with water is a powerful symbol of cleansing and God offers 'streams of living water' to those who believe (John 7:38). In Islam there is washing with water before entering a mosque for prayer. In Hinduism bathing in the sacred Ganges provides a religious cleansing. Many other religions give water an important role in sacred texts and rituals.

Water is important because it underpins our very existence: it is part of our physical, material and spiritual lives. The study of water would therefore also seem to underpin our very existence. Before expanding further on the study of hydrology it is first necessary to step back and take a closer look at the properties of water briefly outlined above. Even though water is the most common substance found on the earth's surface it is also one of the strangest.

Many of these strange properties help to contribute to its importance in sustaining life on earth.

Physical and chemical properties of water

A water molecule consists of two hydrogen atoms bonded to a single oxygen atom (Figure 1.1). The connection between the atoms is through **covalent bonding**: the sharing of an electron from each atom to give a stable pair. This is the strongest type of bonding within molecules and is the reason why water is such a robust compound (i.e. it does not break down into hydrogen and oxygen easily). The robustness of the water molecule means that it stays as a water molecule within our atmosphere because there is not enough energy available to break the covalent bonds and create separate oxygen and hydrogen molecules.

Figure 1.1 shows us that the hydrogen atoms are not arranged around the oxygen atom in a straight line. There is an angle of approximately 105° (i.e. a little larger than a right angle) between the hydrogen atoms. The hydrogen atoms have a positive charge, which means that they repulse each other, but at the same time there are two non-bonding electron pairs on the oxygen atom that also repulse the hydrogen atoms. This leads to the molecular structure shown in Figure 1.1. A water molecule can be described as *bipolar*, which means that there is a positive and negative side to the molecule. This

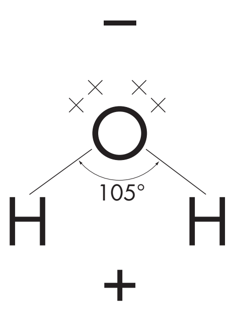

Figure 1.1 The atomic structure of a water molecule. The spare electron pairs on an oxygen atom are shown as small crosses.

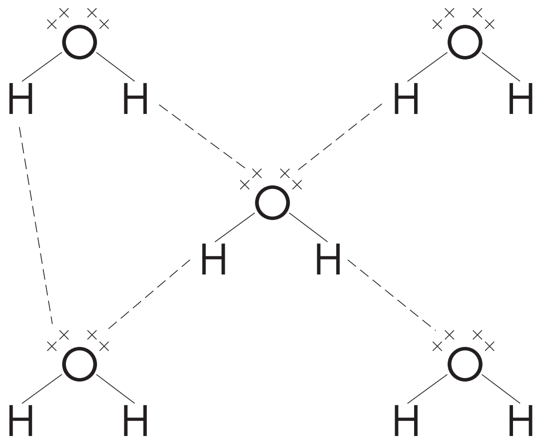

Figure 1.2 The arrangement of water molecules with hydrogen bonds. The stronger covalent bonds between hydrogen and water atoms are shown as solid lines.

Source: Redrawn from McDonald and Kay (1988) and Russell (1976)

polarity is an important property of water as it leads to the bonding between molecules of water: **hydrogen bonding**. The positive side of the molecule (i.e. the hydrogen side) is attracted to the negative side (i.e. the oxygen atom) of another molecule and a weak hydrogen bond is formed (Figure 1.2). The weakness of this bond means that it can be broken with the application of some force and the water molecules separate, forming water in a gaseous state (**water vapour**). Although this sounds easy, it actually takes a lot of energy to break the hydrogen bonds between water molecules. This leads to a high specific heat capacity (see p. 4) whereby a large amount of energy is absorbed by the water to cause a small rise in energy.

The lack of rigidity in the hydrogen bonds between liquid water molecules gives it two more important properties: a low viscosity and the ability to act as an effective solvent. Low viscosity comes from water molecules not being so tightly bound together that they cannot separate when a force is applied to them. This makes water an extremely efficient transport mechanism. When a ship applies force to the water molecules they move aside to let it pass! The ability to act as an efficient solvent comes through water molecules disassociating from each other and being able to surround charged compounds contained within them. As described earlier, the ability of water to act as an efficient solvent allows us to use it for washing, the disposal of pollutants, and also allows nutrients to pass from the soil to a plant.

In water's solid state (i.e. ice) the hydrogen bonds become rigid and a three-dimensional crystalline structure forms. An unusual property of water is that the solid form has a lower density than the liquid form, something that is rare in other compounds. This property has profound implications for the world we live in as it means that ice floats on water. More importantly for aquatic life it means that water freezes from the top down rather the other way around. If water froze from the bottom up, then aquatic flora and fauna would be forced upwards as the water froze and eventually end up stranded on the surface of a pond, river or sea. As it is the flora and fauna are able to survive underneath the ice in liquid water. The maximum density of water actually occurs at around 4°C (see Figure 1.3) so that still bodies of water such as lakes and ponds will display thermal stratification, with water close to 4°C sinking to the bottom.

Water requires a large amount of energy to heat it up. This can be assessed through the **specific heat capacity**, which is the amount of energy required to raise the temperature of a substance by a single degree. Water has a high specific heat capacity relative to other substances (Table 1.1). It requires 4,200 joules of energy to raise the temperature of 1 kilogram of liquid water (approximately 1 litre) by a single degree. In contrast dry soil has a specific heat capacity of around 1.1 kJ/kg/K (it varies according to mineral make up and organic content) and alcohol 0.7 kJ/kg/K. Heating causes the movement of water molecules and that movement requires the breaking of the hydrogen bonds linking them. The large amount of energy required to break the hydrogen bonds in water gives it such a high specific heat capacity.

We can see evidence of water's high specific heat capacity in bathing waters away from the tropics. It is common for sea temperatures to be much lower than air temperatures in high summer since the water is absorbing all the solar radiation and heating up very slowly. In contrast the water temperature also decreases slowly, leading to the sea often being warmer than the air during autumn and winter. As the water cools down it starts to release the energy that it absorbed as it heated up. Consequently for every drop in temperature of 1°C a single kilogram of water releases 4.2 kJ of energy into the atmosphere. It is this that makes water a climate ameliorator. During the summer months a water body will absorb large amounts of energy as it slowly warms up; in an area without a water body, that energy would heat the earth much quicker (i.e. dry soil in Table 1.1) and consequently air temperatures would be higher. In the winter the energy is slowly released from the water as it cools down and is available for heating the atmosphere

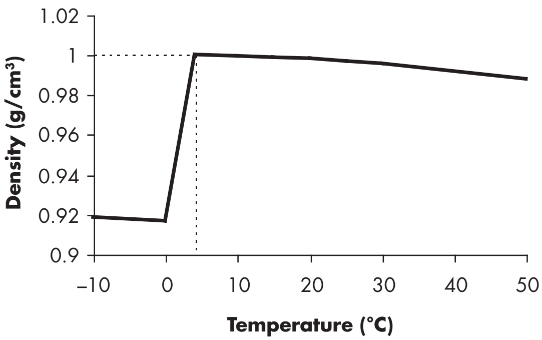

Figure 1.3 The density of water with temperature. The broken line shows the maximum density of water at 3.98°C.

Table 1.1 Specific heat capacity of various substances

Substance	*Specific heat capacity (kJ/kg/K)*
Water	4.2
Dry soil	1.1
Ethanol (alcohol)	0.7
Iron	0.44

nearby. This is why a maritime climate has cooler summers, but warmer winters, than a continental climate.

The energy required to break hydrogen bonds is also the mechanism by which large amounts of energy are transported away from the hot equatorial regions towards the cooler poles. As water evaporates the hydrogen bonds between liquid molecules are broken. This requires a large amount of energy. The first law of thermodynamics states that energy cannot be destroyed, only transformed into another form. In this case the energy absorbed by the water particles while breaking the hydrogen bonds is transformed into latent heat that is then released as sensible heat as the water precipitates (i.e. returns to a liquid form). In the meantime the water has often moved considerable distances in weather systems, taking the latent energy with it. It is estimated that water movement accounts for 70 per cent of lateral global energy transport through latent heat transfer (Mauser and Schädlich, 1998).

Water acts as a climate ameliorator in one other way: water vapour is a powerful greenhouse gas. Radiation direct from the sun (short-wave radiation) passes straight through the atmosphere and may be then absorbed by the earth's surface. This energy is normally re-radiated back from the earth's surface in a different form (long-wave radiation). The long-wave radiation is absorbed by the gaseous water molecules and consequently does not escape the atmosphere. This leads to the gradual warming of the earth–atmosphere system as there is an imbalance between the incoming and outgoing radiation. It is the presence of water vapour in our atmosphere (and other gases such as carbon dioxide and methane) that has allowed the planet to be warm enough to support all of the present life forms that exist.

The catchment or river basin

In studying hydrology the most common spatial unit of consideration is the **catchment** or **river basin**. This can be defined as the area of land from which water flows towards a river and then in that river to the sea. The terminology suggests that the area is analogous to a basin where all water moves towards a central point (i.e. the plug hole, or in this case, the river mouth). The common denominator of any point in a catchment is that wherever rain falls, it will end up in the same place: where the river meets the sea (unless lost through evaporation). A catchment may range in size from a matter of hectares to millions of square kilometres.

A river basin can be defined in terms of its topography through the assumption that all water falling on the surface flows downhill. In this way a catchment boundary can be drawn (as in Figures 1.4 and 1.5) which defines the actual catchment area for a river basin. The assumption that all water flows downhill to the river is not always correct, especially where the underlying geology of a catchment is complicated. It is possible for water to flow as groundwater into another catchment area, creating a problem for the definition of 'catchment area'. These problems aside, the catchment does provide an important spatial unit for hydrologists to consider how water is moving about and is distributed at a certain time.

THE HYDROLOGICAL CYCLE

As a starting point for the study of hydrology it is useful to consider the **hydrological cycle**. This is a conceptual model of how water moves around between the earth and atmosphere in different states as a gas, liquid or solid. As with any conceptual model it contains many gross simplifications; these are discussed in this section. There are different scales that the hydrological cycle can be viewed at, but it is helpful to start at the large global scale and then move to the smaller hydrological unit of a river basin or catchment.

The global hydrological cycle

Table 1.2 sets out an estimate for the amount of water held on the earth at a single time. These figures are extremely hard to estimate accurately.

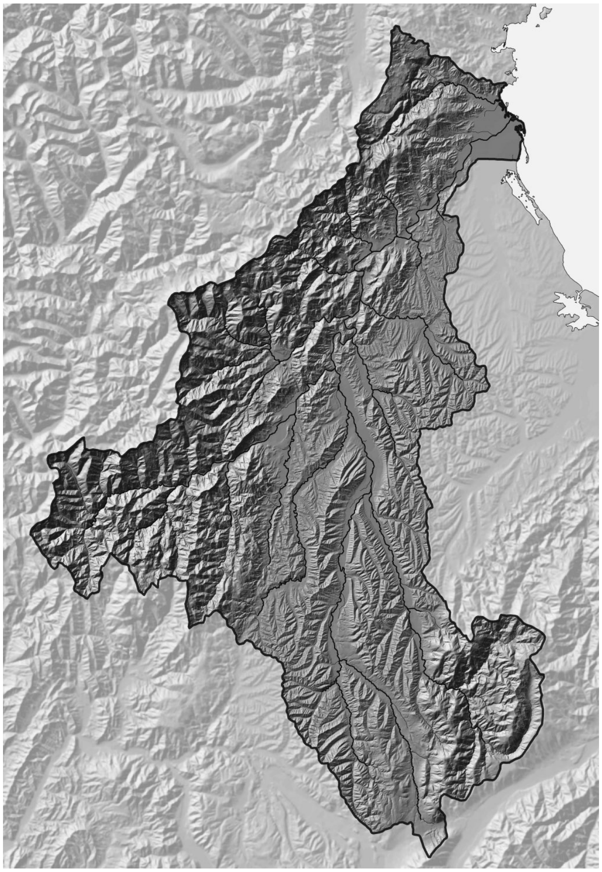

Figure 1.4 (left) Map of the Motueka catchment/watershed, a 2,180 km^2 catchment draining northward at the top of the South Island, New Zealand. Topography is indicated by shading.

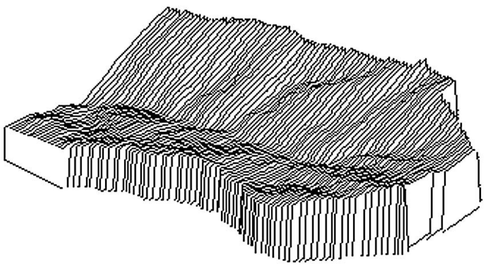

Figure 1.5 A three-dimensional representation of a catchment.

Table 1.2 Estimated volumes of water held at the earth's surface

	Volume ($\times 10^3$ km^3)	*Percentage of total*
Oceans and seas	1,338,000	96.54
Ice caps and glaciers	24,064	1.74
Groundwater	23,400	1.69
Permafrost	300	0.022
Lakes	176	0.013
Soil	16.5	0.001
Atmosphere	12.9	0.0009
Marsh/wetlands	11.5	0.0008
Rivers	2.12	0.00015
Biota	1.12	0.00008
Total	1,385,984	100.00

Source: Data from Shiklomanov and Sokolov (1983)

Estimates cited in Gleick (1993) show a range in total from 1.36 to 1.45 thousand million (or US billion) cubic kilometres of water. The vast majority of this is contained in the oceans and seas. If you were to count groundwater less than 1 km in depth as 'available' and discount snow and ice, then the total percentage of water available for human consumption is around 0.27 per cent. Although this sounds very little it works out at about 146 million litres of water per person per day (assuming a world population of 7 billion); hence the ease with which Stumm (1986) was able to state that there is enough to satisfy all human needs.

Figure 1.6 shows the movement of water around the earth–atmosphere system and is a representation of the global hydrological cycle. The cycle consists of **evaporation** of liquid water into water vapour that is moved around the atmosphere. At some stage the water vapour condenses into a liquid (or solid) again and falls to the surface as **precipitation**. The oceans evaporate more water than they receive as precipitation, while the opposite is true over the continents. The difference between precipitation and evaporation in the terrestrial zone is **runoff**, water moving over or under the surface towards the oceans, which completes the hydrological cycle. As can be seen in Figure 1.6 the vast majority of evaporation and precipitation occurs over the oceans. Ironically this means that the terrestrial zone, which is of greatest concern to hydrologists, is actually rather insignificant in global terms.

The three parts shown in Figure 1.6 (evaporation, precipitation and runoff) are the fundamental processes of concern in hydrology. The figures given in the diagram are global totals but they vary enormously around the globe. This is illustrated in Figure 1.7 which shows how total precipitation is partitioned towards different hydrological processes

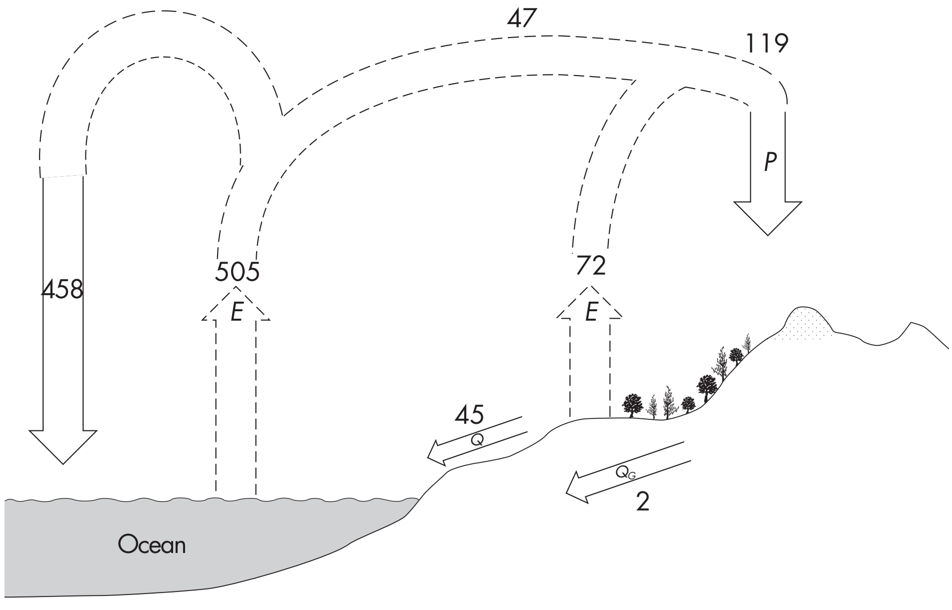

Figure 1.6 The global hydrological cycle. The numbers represent estimates on the total amount of water (thousands of km^3) in each process per annum. E = evaporation; P = precipitation; Q_G = subsurface runoff; Q = surface runoff.

Source: Redrawn from Shiklomanov (1993)

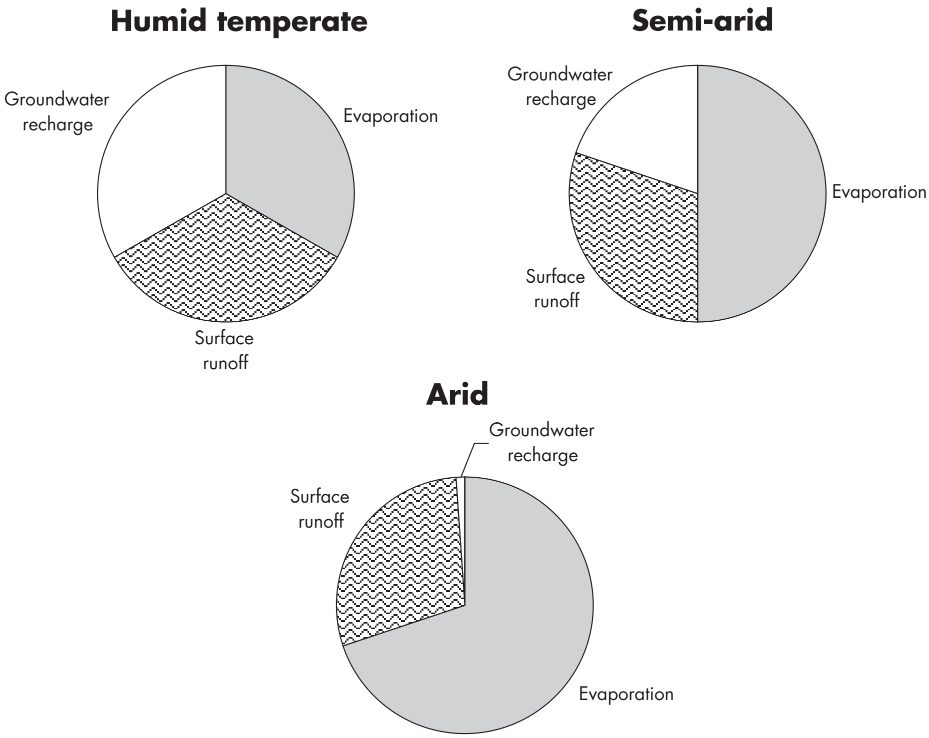

Figure 1.7 Proportion of total precipitation that returns to evaporation, surface runoff or groundwater recharge in three different climate zones.

Source: UNESCO (2006)

in differing amounts depending on climate. In temperate climates (i.e. non tropical or polar) around one third of precipitation becomes evaporation, one third surface runoff and the final third as groundwater recharge. In arid and semi-arid regions the proportion of evaporation is much greater, at the expense of groundwater recharge.

With the advent of satellite monitoring of the earth's surface in the past thirty years it is now possible to gather information on the global distribution of these three processes and hence view how the hydrological cycle varies around the world. In Plates 1 and 2 there are two images of global **rainfall** distribution during 1995, one for January and another for July.

The figure given above of 146 million litres of fresh water per person per year is extremely misleading, as the distribution of available water around the globe varies enormously. The concept of available water considers not only the distribution of rainfall but also population. Table 1.3 gives some indication of those countries that could be considered water rich and water poor in terms of available water. Even this is misleading as a country such as Australia is so large that the high rainfall received in the tropical north-west compensates for the extreme lack of rainfall elsewhere; hence it is considered water rich. The use of rainfall alone is also misleading as it does not consider the importation of water passing across borders, through rivers and groundwater movement.

Table 1.3 gives the amount of available water for various countries, but this takes no account for the amount of water abstracted for actual usage. Figure

1.8 shows the water abstraction per capita for all of the OECD countries. This shows that the USA, Canada and Australia are very high water users, reflecting a very large amount of water used for agricultural and industrial production. The largest water user is the USA with 1,730 m^3 per capita per annum, which is still only 1 per cent of the 146 million litres per capita per annum derived from the Stumm quote. Australia as a high water user has run into enormous difficulties in the years 2005–2007

Table 1.3 Annual renewable water resources per capita (1990 figures) of the seven resource-richest and poorest countries (and other selected countries). Annual renewable water resource is based upon the rainfall within each country; in many cases this is based on estimated figures

Water resource richest countries	*Annual internal renewable water resources per capita (thousand m^3/yr)*	*Water resource poorest countries*	*Annual internal renewable water resources per capita (thousand m^3/yr)*
Iceland	671.9	Bahrain	0.00
Suriname	496.3	Kuwait	0.00
Guyana	231.7	Qatar	0.06
Papua New Guinea	199.7	Malta	0.07
Solomon Islands	149.0	Yemen Arab Republic	0.12
Gabon	140.1	Saudi Arabia	0.16
New Zealand	117.5	United Arab Emirates	0.19
Canada	109.4	Israel	0.37
Australia	20.5	Kenya	0.59
USA	9.9	United Kingdom	2.11

Source: Data from Gleick (1993)

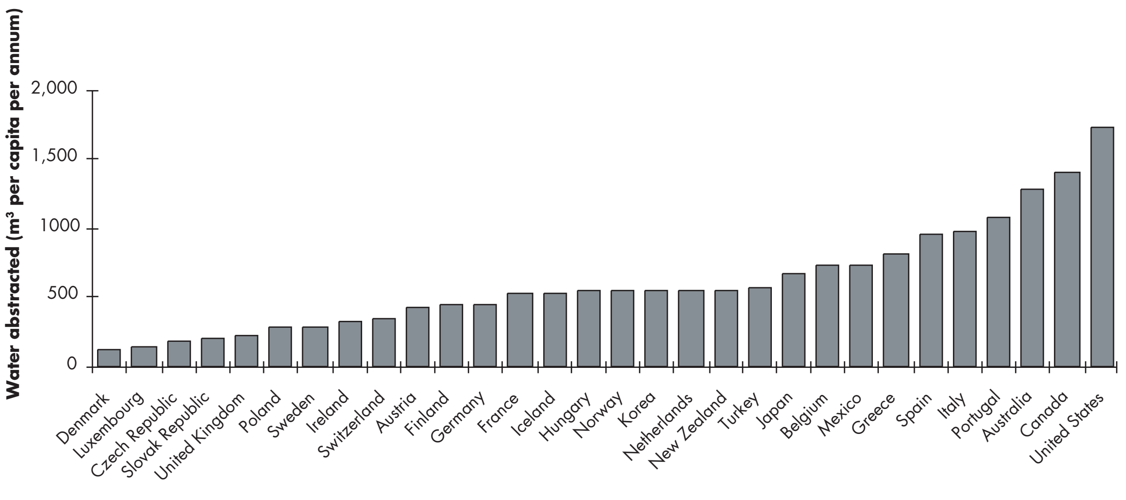

Figure 1.8 Water abstracted per capita for the OECD countries.

Source: OECD Factbook 2005

with severe drought, limiting water availability for domestic and agricultural users. In a situation like this the way that water is allocated (see Chapter 8) literally becomes a matter of life and death, and many economic livelihoods depend on equitable allocation of a scarce water resource.

To try and overcome some of the difficulties in interpreting the data in Figure 1.6 and Table 1.2 hydrologists often work at a scale of more relevance to the physical processes occurring. This is frequently the water basin or catchment scale (Figures 1.4 and 1.5).

The catchment hydrological cycle

At a smaller scale it is possible to view the catchment hydrological cycle as a more in-depth conceptual model of the hydrological processes operating. Figure 1.9 shows an adaptation of the global hydrological cycle to show the processes operating within a catchment. In Figure 1.9 there are still essentially three processes operating (evaporation, precipitation and runoff), but it is possible to subdivide each into different sub-processes. Evaporation is a mixture of open water evaporation (i.e. from rivers and lakes); evaporation from the soil; evaporation from plant surfaces; **interception**; and **transpiration** from plants. Precipitation can be in the form of **snowfall**, hail, rainfall or some mixture of the three (sleet). Interception of precipitation by plants makes the water available for evaporation again before it even reaches the soil surface. The broad term 'runoff' incorporates the movement of liquid water above and below the surface of the earth. The movement of water below the surface necessitates an understanding of infiltration into the soil and how the water moves in the unsaturated zone (**throughflow**) and in the saturated zone (**groundwater flow**). All of these processes and sub-processes are dealt with in detail in later chapters; what is important to realise at this stage is that it is part of one continuous cycle that moves water around the globe and that they may all be operating at different times within a river basin.

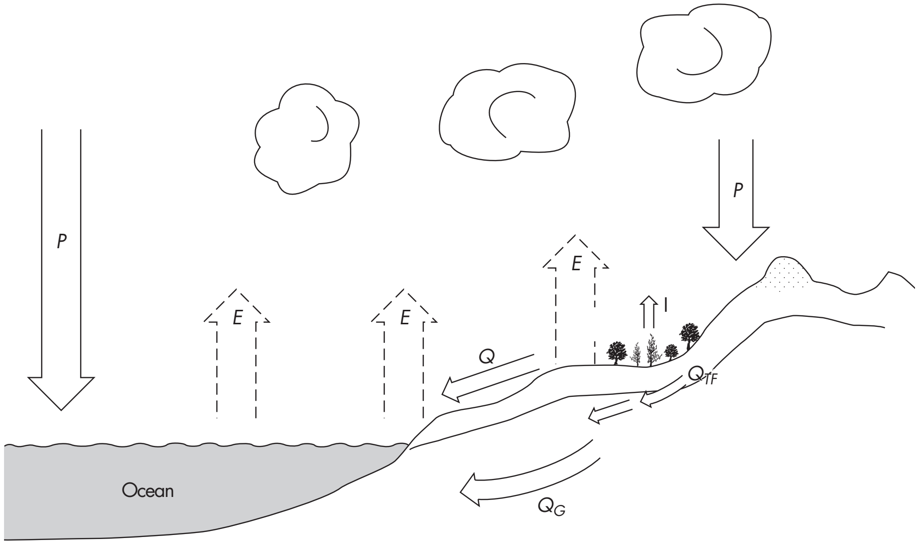

Figure 1.9 Processes in the hydrological cycle operating at the basin or catchment scale. Q = runoff; the subscript G stands for groundwater flow; TF for throughflow; I = interception; E = evaporation; P = precipitation.

THE WATER BALANCE EQUATION

In the previous section it was stated that the hydrological cycle is a conceptual model representing our understanding of which processes are operating within an overall earth–atmosphere system. It is also possible to represent this in the form of an equation, which is normally termed the **water balance equation**. The water balance equation is a mathematical description of the hydrological processes operating within a given timeframe and incorporates principles of mass and energy continuity. In this way the hydrological cycle is defined as a closed system whereby there is no mass or energy created or lost within it. The mass of concern in this case is water.

There are numerous ways of representing the water balance equation but equation 1.1 shows it in its most fundamental form.

$$P \pm E \pm \Delta S \pm Q = 0 \tag{1.1}$$

where P is precipitation; E is evaporation; ΔS is the change in **storage** and Q is runoff. Runoff is normally given the notation of Q to distinguish it from rainfall which is often given the symbol R and frequently forms the major component of precipitation. The ± terminology in equation 1.1 represents the fact that each term can be either positive or negative depending on which way you view it – for example, precipitation is a gain (positive) to the earth but a loss (negative) to the atmosphere. As most hydrology is concerned with water on or about the earth's surface it is customary to consider the terms as positive when they represent a gain to the earth.

Two of the more common ways of expressing the water balance are shown in equations 1.2 and 1.3

$$P - Q - E - \Delta S = 0 \tag{1.2}$$

$$Q = P - E - \Delta S \tag{1.3}$$

In equations 1.2 and 1.3 the change in storage term can be either positive or negative, as water can be released from storage (negative) or absorbed into storage (positive).

The terms in the water balance equation can be recognised as a series of fluxes and stores. A **flux** is a rate of flow of some quantity (Goudie *et al*., 1994): in the case of hydrology the quantity is water. The water balance equation assesses the relative flux of water to and from the surface with a storage term also incorporated. A large part of hydrology is involved in measuring or estimating the amount of water involved in this flux transfer and storage of water.

Precipitation in the water balance equation represents the main input of water to a surface (e.g. a catchment). As explained on p. 10, precipitation is a flux of both rainfall and snowfall. Evaporation as a flux includes that from open water bodies (lakes, ponds, rivers), the soil surface and vegetation (including both interception and transpiration from plants). The storage term includes soil moisture, deep groundwater, water in lakes, glaciers, seasonal snow cover. The runoff flux is also explained on p. 10. In essence it is the movement of liquid water above and below the surface of the earth.

The water balance equation is probably the closest that hydrology comes to having a fundamental theory underlying it as a science, and hence almost all hydrological study is based around it. Field catchment studies are frequently trying to measure the different components of the equation in order to assess others. Nearly all hydrological **models** attempt to solve the equation for a given time span – for example, by knowing the amount of rainfall for a given area and estimating the amount of evaporation and change in storage it is possible to calculate the amount of runoff that might be expected.

Despite its position as a fundamental hydrological theory there is still considerable uncertainty about the application of the water balance equation. It is not an uncertainty about the equation itself but rather about how it may be applied. The problem is that all of the processes occur at a spatial and temporal scale (i.e. they operate over a period of time and within a certain area) that may not coincide with the scale at which we make our measurement or estimation. It is this issue of *scale* that makes hydrology appear an imprecise science and it will be discussed further in the remaining chapters of this book.

OUTLINE OF THE BOOK

Fundamentals of Hydrology attempts to bring out the underlying principles in the science of hydrology and place these in a water management context. By and large, water management is concerned with issues of water quantity (floods, droughts, water distribution . . .) and water quality (drinking water, managing aquatic ecosystems . . .). These two management concerns forms the basis for discussion within the book. It starts with the four components of the water balance equation (i.e. precipitation, evaporation, change in storage and runoff) in Chapters 2–5. Precipitation is dealt with in Chapter 2, followed by evaporation, including canopy interception, in Chapter 3. Chapter 4 looks at the storage term from the water balance equation, in particular the role of water stored under the earth's surface as soil water and groundwater and also storage as snow and ice. Chapter 5 is concerned with the runoff processes that lead to water flowing down a channel in a stream or river.

Each of Chapters 2–5 starts with a detailed description of the process under review in the chapter. They then move on to contain a section on how it is possible to measure the process, followed by a section on how it may be estimated. In reality it is not always possible to separate between measurement and estimation as many techniques contain an element of both within them, something that is pointed out in various places within these chapters. Chapters 2–5 finish with a discussion on how the particular process described has relevance to water quantity and quality.

Chapter 6 moves away from a description of process and looks at the methods available to analyse streamflow records. This is one of the main tasks within hydrology and three particular techniques are described: hydrograph analysis (including the unit hydrograph), flow duration curves and frequency analysis. The latter mostly concentrates on **flood frequency analysis**, although there is a short description of how the techniques can be applied to low flows. The chapter also has sections on hydrological modelling and combining ecology and hydrology for instream flow analysis.

Chapter 7 is concerned with water quality in the fresh water environment. This chapter has a description of major water quality parameters, measurement techniques and some strategies used to control water quality.

The final chapter takes an integrated approach to look at different issues of change that affect hydrology. This ranges from water resource management and a changing legislative framework to climate and land use change. These issues are discussed with reference to research studies investigating the different themes. It is intended as a way of capping off the fundamentals of hydrology by looking at real issues facing hydrology in the twenty-first century.

ESSAY QUESTIONS

1 **Discuss the nature of water's physical properties and how important these are in determining the natural climate of the earth.**
2 **Describe how the hydrological cycle varies around the globe.**
3 **How may water-poor countries overcome the lack of water resources within their borders?**

WEBSITES

A warning: although it is often easy to access information via the World Wide Web you should always be careful in utilising it. There is no control on the type of information available or on the data presented. More traditional channels, such as research journals and books, undergo a peer review process where there is some checking of content. This may happen for websites but there is no guarantee that it has happened. You should be wary of treating everything read from the World Wide Web as being correct.

The websites listed here are general sources of hydrological information that may enhance the reading of this book. The majority of addresses are included for the web links provided within their sites. The web addresses were up to date in early 2007 but may change in the future. Hopefully there is enough information provided to enable the use of a search engine to locate updated addresses.

http://www.cig.ensmp.fr/~iahs
International Association of Hydrological Sciences (IAHS): a constituent body of the International Union of Geodesy and Geophysics (IUGG), promoting the interests of hydrology around the world. This has a useful links page.

http://www.cig.ensmp.fr/~hubert/glu/aglo.htm
Part of the IAHS site, this provides a glossary of hydrological terms (in multiple languages).

http://www.worldwater.org
The World's Water, part of the Pacific Institute for Studies in Development, Environment, and Security: this is an organisation that studies water resource issues around the world. There are some useful information sets here.

http://www.ucowr.siu.edu/
Universities Council on Water Resources: 'universities and organizations leading in education, research and public service in water resources'. Disseminates information of interest to the water resources community in the USA.

http://www.ewatercrc.com.au/
Ewater is the successor to the previous Cooperative Research Centre for Catchment Hydrology: an Australian research initiative that focuses on tools and information of use in catchment management.

http://www.wsag.unh.edu/
Water Systems Analysis Group at the University of New Hampshire: undertakes a diverse group of hydrological research projects at different scales and regions. Much useful information and many useful links.

http://www.hydrologynz.org.nz/
Home site for the New Zealand Hydrological Society: has a links page with many hydrological links.

http://www.cof.orst.edu/cof/fe/watershed
Hillslope and Watershed Hydrology Team at Oregon State University: this has many good links and information on the latest research.

http://www.ceh.ac.uk/
Centre for Ecology and Hydrology (formerly Institute of Hydrology) in the UK: a hydrological research institute. There is a very good worldwide links page here.

http://water.usgs.gov
Water Resources Division of the United States Geological Survey (USGS): provides information on groundwater, surface water and water quality throughout the USA.

http://ghrc.msfc.nasa.gov
Global Hydrology Resource Centre: a NASA site with mainly remote sensing data sets of relevance for global hydrology.

http://www.whycos.org/
WHYCOS is a World Meteorological Organization (WMO) programme aiming at improving the basic observation activities, strengthening international cooperation and promoting free exchange of data in the field of hydrology.This website provides information on the System, projects, technical materials, data and links.

http://www.who.int/water_sanitation_health/diseases/en/
This World Health Organization (WHO) section contains fact sheets on over twenty water-related diseases, estimates of the global burden of water-related disease, information on water requirements (quantity, service level) to secure health benefits, and facts and figures on water, sanitation and hygiene links to health.

http://www.unesco.org/water/water_links/
A comprehensive set of hydrological links that can be searched under different themes (e.g. droughts, floods), geographic regions or organisations.

2

PRECIPITATION

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the processes of precipitation formation.
- A knowledge of the techniques for measuring precipitation (rainfall and snow).
- An appreciation of the associated errors in measuring precipitation.
- A knowledge of how to analyse rainfall data spatially and for intensity/duration of a storm.
- A knowledge of some of the methods used to estimate rainfall at the large scale.
- An understanding of the process of precipitation interception by a canopy.

PRECIPITATION AS A PROCESS

Precipitation is the release of water from the atmosphere to reach the surface of the earth. The term 'precipitation' covers all forms of water being released by the atmosphere, including snow, hail, sleet and rainfall. It is the major input of water to a river catchment area and as such needs careful assessment in any hydrological study. Although rainfall is relatively straightforward to measure (other forms of precipitation are more difficult) it is notoriously difficult to measure *accurately* and, to compound the problem, is also extremely variable within a catchment area.

Precipitation formation

The ability of air to hold water vapour is temperature dependent: the cooler the air the less water vapour is retained. If a body of warm, moist air is cooled then it will become saturated with water vapour and eventually the water vapour will condense into liquid or solid water (i.e. water or ice droplets). The water will not condense spontaneously however; there need to be minute particles present in the atmosphere, called **condensation nuclei**, upon which the water or ice droplets form. The water or ice droplets that form on condensation nuclei are normally too small to fall to the surface as precipitation; they need to grow in order to have enough mass to overcome uplifting forces within a cloud. So there are three conditions that need to be met prior to precipitation forming:

1 Cooling of the atmosphere
2 Condensation onto nuclei
3 Growth of the water/ice droplets.

Atmospheric cooling

Cooling of the atmosphere may take place through several different mechanisms occurring independently or simultaneously. The most common form of cooling is from the uplift of air through the atmosphere. As air rises the pressure decreases; Boyle's Law states that this will lead to a corresponding cooling in temperature. The cooler temperature leads to less water vapour being retained by the air and conditions becoming favourable for **condensation**. The actual uplift of air may be caused by heating from the earth's surface (leading to **convective precipitation**), an air mass being forced to rise over an obstruction such as a mountain range (this leads to **orographic precipitation**), or from a low pressure weather system where the air is constantly being forced upwards (this leads to **cyclonic precipitation**). Other mechanisms whereby the atmosphere cools include a warm air mass meeting a cooler air mass, and the warm air meeting a cooler object such as the sea or land.

Condensation nuclei

Condensation nuclei are minute particles floating in the atmosphere which provide a surface for the water vapour to condense into liquid water upon. They are commonly less than a micron (i.e. one-millionth of a metre) in diameter. There are many different substances that make condensation nuclei, including small dust particles, sea salts and smoke particles.

Research into generating artificial rainfall has concentrated on the provision of condensation nuclei into clouds, a technique called **cloud seeding**. During the 1950s and 1960s much effort was expended in using silver iodide particles, dropped from planes, to act as condensation nuclei. However, more recent work has suggested that other salts such as potassium chloride are better nuclei. There is much controversy over the value of cloud seeding. Some studies support its effectiveness (e.g. Gagin and Neumann, 1981; Ben-Zvi, 1988); other authors query the results (e.g. Rangno and Hobbs, 1995), while others suggest that it only works in certain atmospheric conditions and with certain cloud types (e.g. Changnon *et al*., 1995). More recent work in South Africa has concentrated on using hygroscopic flares to release chloride salts into the base of convective storms, with some success (Mather *et al*., 1997). Interestingly, this approach was first noticed through the discovery of extra heavy rainfall occurring over a paper mill in South Africa that was emitting potassium chloride from its chimney stack (Mather, 1991).

Water droplet growth

Water or ice droplets formed around condensation nuclei are normally too small to fall directly to the ground; that is, the forces from the upward draught within a cloud are greater than the gravitational forces pulling the microscopic droplet downwards. In order to overcome the upward draughts it is necessary for the droplets to grow from an initial size of 1 micron to around 3,000 microns (3 mm). The vapour pressure difference between a droplet and the surrounding air will cause it to grow through condensation, albeit rather slowly. When the water droplet is ice the vapour pressure difference with the surrounding air becomes greater and the water vapour sublimates onto the ice droplet. This will create a precipitation droplet faster than condensation onto a water droplet, but is still a slow process. The main mechanism by which raindrops grow within a cloud is through *collision and coalescence*. Two raindrops collide and join together (coalesce) to form a larger droplet that may then collide with many more before falling towards the surface as rainfall or another form of precipitation.

Another mechanism leading to increased water droplet size is the so-called **Bergeron process**. The pressure exerted within the parcel of air, by having the water vapour present within it, is called the **vapour pressure**. The more water vapour present the greater the vapour pressure. Because there is a maximum amount of water vapour that can be held by the parcel of air there is also a maximum vapour pressure, the so-called **saturation vapour pressure**. The saturation vapour pressure is greater over a water droplet than an ice droplet because it is easier for water molecules to escape from the surface of a liquid than a solid. This creates a water vapour gradient between water droplets and ice crystals so that water vapour moves from the water droplets to the ice crystals, thereby increasing the size of the ice crystals. Because clouds are usually a mixture of water vapour, water droplets and ice crystals, the Bergeron process may be a significant factor in making water droplets large enough to become rain drops (or ice/snow crystals) that overcome gravity and fall out of the clouds.

The mechanisms of droplet formation within a cloud are not completely understood. The relative proportion of condensation-formed, collision-formed, and Bergeron-process-formed droplets depends very much on the individual cloud circumstances and can vary considerably. As a droplet is moved around a cloud it may freeze and thaw several times, leading to different types of precipitation (see Table 2.1).

Dewfall

The same process of condensation occurs in **dewfall**, only in this case the water vapour condenses into liquid water after coming into contact with a cold surface. In humid-temperate countries dew is a common occurrence in autumn when the air at night is still warm but vegetation and other surfaces have cooled to the point where water vapour coming into contact with them condenses onto the leaves and forms dew. Dew is not normally a major part of the hydrological cycle but is another form of precipitation.

PRECIPITATION DISTRIBUTION

The amount of precipitation falling over a location varies both spatially and temporally (with time). The different influences on the precipitation can be divided into static and dynamic influences. Static influences are those such as altitude, aspect and slope; they do not vary between storm events. Dynamic influences are those that do change and are by and large caused by variations in the weather. At the global scale the influences on precipitation distribution are mainly dynamic being caused by differing weather patterns, but there are static factors such as topography that can also cause major variations through a **rain shadow effect** (see case study on pp. 18–19). At the continental scale large differences in rainfall can be attributed to a mixture

Table 2.1 Classes of precipitation used by the UK Meteorological Office

Class	*Definition*
Rain	Liquid water droplets between 0.5 and 7 mm in diameter
Drizzle	A subset of rain with droplets less than 0.5 mm
Sleet	Freezing raindrops; a combination of snow and rain
Snow	Complex ice crystals agglomerated
Hail	Balls of ice between 5 and 125 mm in diameter

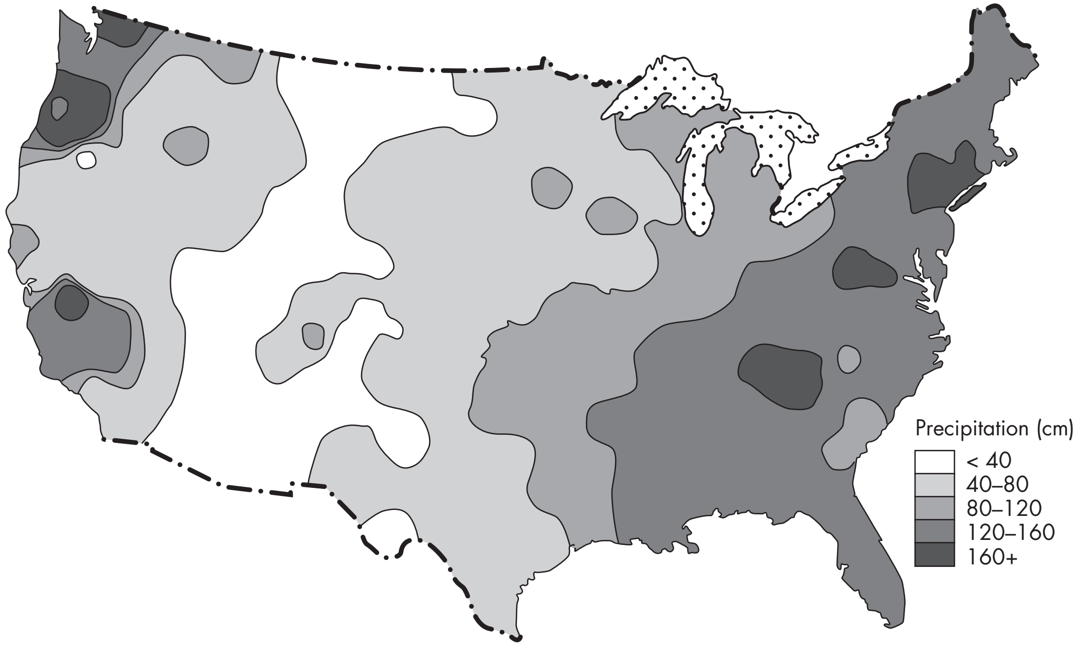

Figure 2.1 Annual precipitation across the USA during 1996.

Source: Redrawn with data from the National Atmospheric Deposition Program

of static and dynamic factors. In Figure 2.1 the rainfall distribution across the USA shows marked variations. Although mountainous areas have a higher rainfall, and also act as a block to rainfall reaching the drier centre of the country, they do not provide the only explanation for the variations evident in Figure 2.1. The higher rainfall in the north-west states (Oregon and Washington) is linked to wetter cyclonic weather systems from the northern Pacific that do not reach down to southern California. Higher rainfall in Florida and other southern states is linked to the warm waters of the Caribbean sea. These are examples of dynamic influences as they vary between rainfall events.

At smaller scales the static factors are often more dominant, although it is not uncommon for quite large variations in rainfall across a small area caused by individual storm clouds to exist. As an example: on 3 July 2000 an intense rainfall event caused flooding in the village of Epping Green, Essex, UK. Approximately 10 mm of rain fell within one hour, although there was no recorded rainfall in the village of Theydon Bois approximately 10 km to the south. This large spatial difference in rainfall was caused by the scale of the weather system causing the storm – in this case a convective thunderstorm. Often these types of variation lessen in importance over a longer timescale so that the annual rainfall in Epping Green and Theydon Bois is very similar, whereas the daily rainfall may differ considerably. For the hydrologist, who is interested in rainfall at the small scale, the only way to try and characterise these dynamic variations is through having as many **rain gauges** as possible within a study area.

Static influences on precipitation distribution

It is easier for the hydrologist to account for static variables such as those discussed below.

Altitude

It has already been explained that temperature is a critical factor in controlling the amount of water vapour that can be held by air. The cooler the air is, the less water vapour can be held. As temperature decreases with altitude it is reasonable to assume that as an air parcel gains altitude it is more likely to release the water vapour and cause higher rainfall. This is exactly what does happen and there is a strong correlation between altitude and rainfall: so-called *orographic precipitation*.

Aspect

The influence of aspect is less important than altitude but it may still play an important part in the distribution of precipitation throughout a catchment. In the humid mid-latitudes (35° to 65° north or south of the equator) the predominant source of rainfall is through cyclonic weather systems arriving from the west. Slopes within a catchment that face eastwards will naturally be more sheltered from the rain than those facing westwards. The same principle applies everywhere: slopes with aspects facing away from the predominant weather patterns will receive less rainfall than their opposites.

Slope

The influence of slope is only relevant at a very small scale. Unfortunately the measurement of rainfall occurs at a very small scale (i.e. a rain gauge). The difference between a level rain gauge on a hillslope, compared to one parallel to the slope, may be significant. It is possible to calculate this difference if it is assumed that rain falls vertically – but of course rain does not always fall vertically. Consequently the effect of slope on rainfall measurements is normally ignored.

Rain shadow effect

Where there is a large and high land mass it is common to find the rainfall considerably higher on one side than the other. This is through a combination of altitude, slope, aspect and dynamic weather direction influences and can occur at many different scales (see Case Study below).

Case study

THE RAIN SHADOW EFFECT

The predominant weather pattern for the South Island of New Zealand is a series of rain-bearing depressions sweeping up from the Southern Ocean, interrupted by drier blocking anticyclones. The Southern Alps form a major barrier to the fast-moving depressions and have a huge influence on the rainfall distribution within the South Island. Formed as part of tectonic uplift along the Pacific/Indian plate boundary, the Southern Alps stretch the full length of the South Island (approximately 700 km) and at their highest point are over 3,000 m above mean sea level.

The predominant weather pattern has a westerly airflow, bringing moist air from the Tasman Sea onto the South Island. As this air is forced up over the Southern Alps it cools down and releases much of its moisture as rain and snow. As the air descends on the eastern side of the mountains it warms up and becomes a föhn wind, referred to locally as a ‘nor-wester’. The annual rainfall patterns for selected stations in the South Island are shown in Figure 2.2. The rain shadow effect can be clearly seen with the west coast rainfall being at least four times that of the east. Table 2.2

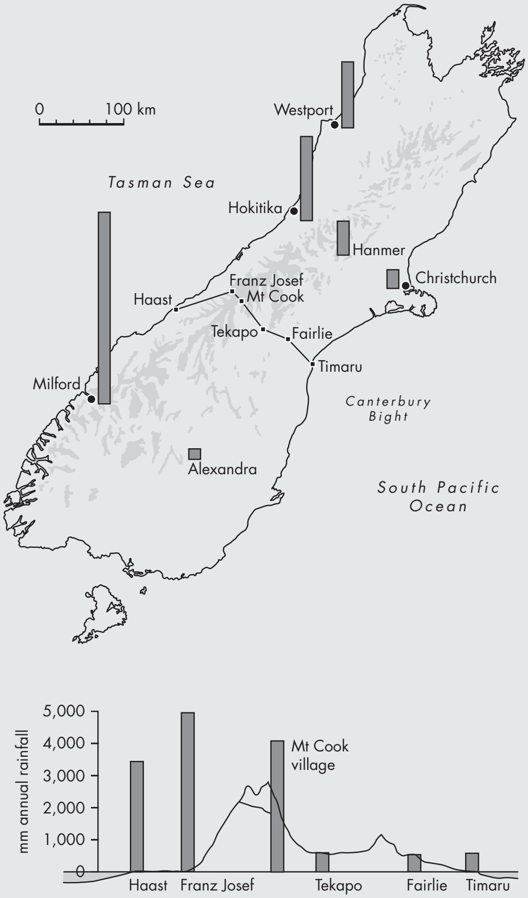

Figure 2.2 Rainfall distribution across the Southern Alps of New Zealand (South Island). Shaded areas on the map are greater than 1,500 m in elevation. A clear rain shadow effect can be seen between the much wetter west coast and the drier east.

also illustrates the point, with the number of rain days at different sites in a cross section across the South Island. Although not shown on the transect in Figure 2.2 recordings of rainfall further north in the Southern Alps (Cropp River inland from Hokitika) are as high as 6 m a year.

This pattern of rain shadow is seen at many different locations around the globe. It does not require as large a barrier as the Southern Alps – anywhere with a significant topographical barrier is likely to cause some form of rain shadow. Hayward and Clarke (1996) present data showing a strong rain shadow across the Freetown Peninsula in Sierra Leone. They analysed mean monthly rainfall in 31 gauges within a 20 × 50 km area, and found that the rain shadow effect was most marked during the monsoon months of June to October. The gauges in locations facing the ocean (south-west aspect) caught considerably more rainfall during the monsoon than those whose aspect was towards the north-east and behind a small range of hills.

Table 2.2 Average annual rainfall and rain days for a cross section across the South Island

Weather station	*Height above mean sea level*	*Annual rainfall (mm)*	*Rain days*
Haast	30	5,840	175
Mt Cook village	770	670	120
Tekapo	762	604	77
Timaru	25	541	75

Note: More details on weather differentials across the South Island of New Zealand are in Sinclair *et al.* (1996)
Source: Data from New Zealand Met. Service and other miscellaneous sources

Forest rainfall partitioning

Once rain falls onto a vegetation canopy it effectively partitions the water into separate modes of movement: **throughfall**, **stemflow** and **interception loss**. This is illustrated in Figure 2.3.

Throughfall

This is the water that falls to the ground either directly, through gaps in the canopy, or indirectly, having dripped off leaves, stems or branches. The amount of *direct throughfall* is controlled by the canopy coverage for an area, a measure of which is the leaf area index (LAI). LAI is actually the ratio of leaf area to ground surface area and consequently has a value greater than one when there is more than one layer of leaf above the ground. When the LAI is less than one you would expect some direct throughfall to occur. When you shelter under a tree during a rainstorm you are trying to avoid the rainfall and direct throughfall. The greater the surface area of leaves above you, the more likely it is that you will avoid getting wet from direct throughfall.

The amount of *indirect throughfall* is also controlled by the LAI, in addition to the **canopy storage capacity** and the rainfall characteristics. Canopy storage capacity is the volume of water that can be held by the canopy before water starts dripping as indirect throughfall. The canopy storage capacity is controlled by the size of trees, plus the area and water-holding capacity of individual leaves. Rainfall characteristics are an important control on indirect throughfall as they dictate how quickly the canopy storage capacity is filled. Experience of standing under trees during a rainstorm should tell you that intensive rainfall quickly turns into indirect throughfall (i.e. you get wet!), whereas light showers frequently do not reach the ground surface at all. In reality canopy storage capacity is a rather nebulous concept. Canopy characteristics are constantly changing and it is rare for water on a canopy to fill up completely before creating indirect

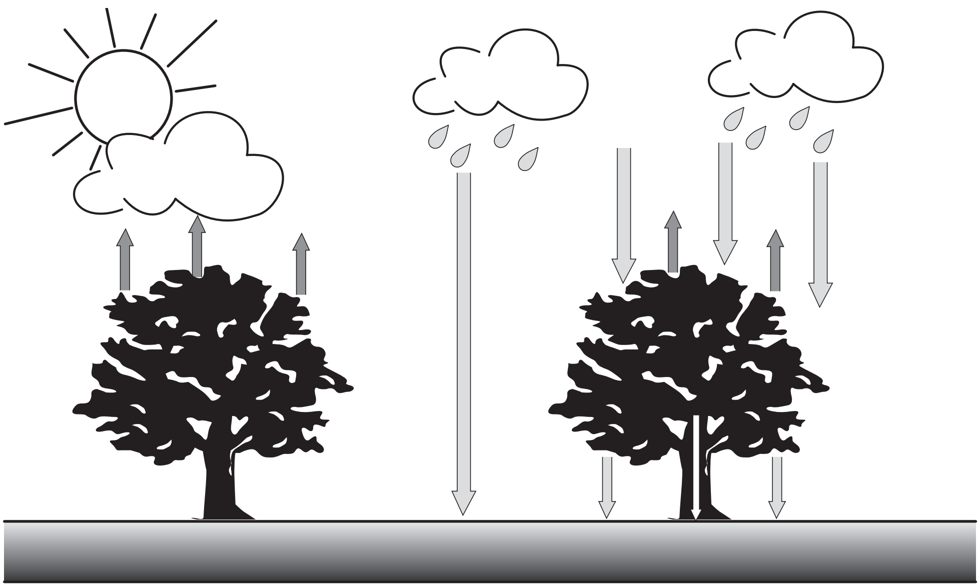

Figure 2.3 Rainfall above and below a canopy. Indicated on the diagram are stemflow (white arrow on trunk); direct and indirect throughfall (lightly hatched arrow); and interception loss (upward-facing darker arrow).

throughfall. This means that indirect throughfall occurs before the amount of rainfall equals the canopy storage capacity, making it difficult to gauge exactly what the storage capacity is.

Stemflow

Stemflow is the rainfall that is intercepted by stems and branches and flows down the tree trunk into the soil. Although measurements of stemflow show that it is a small part of the hydrological cycle (normally 2–10 per cent of above canopy rainfall; Lee, 1980) it can have a much more significant role. Durocher (1990) found that trees with smoother bark such as beech (*Fagus*) had higher rates of stemflow as the smoothness of bark tends to enhance drainage towards stemflow.

Stemflow acts like a funnel (see Figure 2.4), collecting water from a large area of canopy but delivering it to the soil in a much smaller area: the surface of the trunk at the base of a tree. This is most obvious for the deciduous oak-like tree illustrated in Figure 2.4, but it still applies for other structures (e.g. conifers) where the area of stemflow entry into the soil is far smaller than the canopy catchment area for rainfall. At the base of a tree it is possible for the water to rapidly enter the soil through flow along roots and other macropores surrounding the root structure. This can act as a rapid conduit of water sending a significant pulse into the **soil water**.

Interception loss

While water sits on the canopy, prior to indirect throughfall or stemflow, it is available for evaporation, referred to as *interception loss*. This is an evaporation process and it is discussed further in the following chapter.

Interception gain

In some circumstances it is possible that there is an interception gain from vegetation. In the Bull Run catchment, Oregon, USA it has been shown that the water yield after timber harvesting was significantly less than prior to the trees being logged (Harr, 1982; Ingwersen, 1985). This is counter to the majority of catchment studies reported by Bosch and Hewlett (1982) which show an increase in water yield as forests are logged. The reason for the loss of water with the corresponding loss of trees in Oregon

Figure 2.4 The funnelling effect of a tree canopy on stemflow.

is to do with the particular circumstances of the catchment. Fog from the cold North Pacific, with no accompanying rain, is a common feature and it is believed that the trees intercept fog particles, creating ‘fog drip’ which is a significant part of the water balance. Fog droplets are extremely small and Ingwersen (1985) has suggested that the sharp ends of needles on pine trees act as condensation nuclei, promoting the growth of larger droplets that fall to the ground (see an example of fogdrip from tussock leaves in Plate 3). When the trees are removed there are no condensation nuclei (or far fewer) on the resultant vegetation so the water remains in the atmosphere and is ‘lost’ in terms of water yield. Equally important is the influence of vegetation roughness. The turbulent mixing of air as wind passes over a rough canopy promotes rapid deposition of condensing water (directly converse to interception loss, see Chapter 3). The overall result of this is that the removal of trees leads to less water in the river; this runs counter to the evidence provided in the Case Study in Chapter 4.

MEASUREMENT

For hydrological analysis it is important to know how much precipitation has fallen and when this occurred. The usual expression of precipitation is as a vertical depth of liquid water. Rainfall is measured by millimetres or inches depth, rather than by volume such as litres or cubic metres. The measurement is the depth of water that would accumulate on the surface if all the rain remained where it had fallen (Shaw, 1994). Snowfall may also be expressed as a depth, although for hydrological purposes it is most usefully described in water equivalent depth (i.e. the depth of water that would be present if the snow melted). This is a recognition that snow takes up a greater volume (as much as 90 per cent more) for the same amount of liquid water.

There is a strong argument that can be made to say that there is no such thing as precipitation measurement at the catchment scale as it varies so tremendously over a small area. The logical end-point to this argument is that all measurement techniques are in fact precipitation estimation techniques. For the sake of clarity in this text precipitation measurement techniques refer to the methods used to quantify the volume of water present, as opposed to estimation techniques where another variable is used as a surrogate for the water volume.

Rainfall measurement

The instrument for measuring rainfall is called a *rain gauge*. A rain gauge measures the volume of water that falls onto a horizontal surface delineated by the rain gauge rim (see Figure 2.5). The volume is converted into a rainfall depth through division by the rain gauge surface area. The design of a rain gauge is not as simple as it may seem at first glance as there are many errors and inaccuracies that need to be minimised or eliminated.

Figure 2.5 A rain gauge sitting above the surface to avoid splash.

There is a considerable scientific literature studying the accuracy and errors involved in measuring rainfall. It needs to be borne in mind that a rain gauge represents a very small point measurement (or sample) from a much larger area that is covered by the rainfall. Any errors in measurement will be amplified hugely because the rain gauge collection area represents such a small sample size. Because of this amplification it is extremely important that the design of a rain gauge negates any errors and inaccuracies.

The four main sources of error in measuring rainfall that need consideration in designing a method for the accurate measurement of rainfall are:

1 Losses due to evaporation
2 Losses due to wetting of the gauge
3 Over-measurement due to splash from the surrounding area
4 Under-measurement due to turbulence around the gauge.

Evaporation losses

A rain gauge can be any collector of rainfall with a known collection area; however, it is important that any rainfall that does collect is not lost again through evaporation. In order to eliminate, or at least lessen this loss, rain gauges are funnel shaped. In this way the rainfall is collected over a reasonably large area and then any water collected is passed through a narrow aperture to a collection tank underneath. Because the collection tank has a narrow top (i.e. the funnel mouth) there is very little interchange of air with the atmosphere above the gauge. As will be explained in Chapter 3, one of the necessary requirements for evaporation is the turbulent mixing of saturated air with drier air above. By restricting this turbulent transfer there is little evaporation that can take place. In addition to this, the water awaiting measurement is kept out of direct sunlight so that it will not be warmed; hence there is a low evaporation loss.

Wetting loss

As the water trickles down the funnel it is inevitable that some water will stay on the surface of the funnel and can be lost to evaporation or not measured in the collection tank. This is often referred to as a *wetting loss*. These losses will not be large but may be significant, particularly if the rain is falling as a series of small events on a warm day. In order to lessen this loss it is necessary to have steep sides on the funnel and to have a non-stick surface. The standard UK Meteorological Office rain gauge is made of copper to create a non-stick surface, although many modern rain gauges are made of non-adhesive plastics.

Rain splash

The perfect rain gauge should measure the amount of rainfall that would have fallen on a surface if the gauge was not there. This suggests that the ideal situation for a rain gauge is flush with the surface. A difficulty arises, however, as a surface-level gauge is likely to over-measure the catch due to rain landing adjacent to the gauge and splashing into it. If there was an equal amount of splash going out of the gauge then the problem might not be so severe, but the sloping sides of the funnel (to reduce evaporative losses) mean that there will be very little splash-out. In extreme situations it is even possible that the rain gauge could be flooded by water flowing over the surface or covered by snow. To overcome the splash, flooding and snow coverage problem the rain gauge can be raised up above the ground (Figure 2.5) or placed in the middle of a non-splash grid (see Figure 2.6).

Turbulence around a raised gauge

If a rain gauge is raised up above the ground (to reduce splash) another problem is created due to air turbulence around the gauge. The rain gauge presents an obstacle to the wind and the consequent aerodynamic interference leads to a reduced catch (see Figure 2.7). The amount of loss is dependent on

Figure 2.6 Surface rain gauge with non-splash surround.

Figure 2.8 Baffles surrounding a rain gauge to lessen the impact of wind turbulence. The gauge is above ground because of snow cover during the winter.

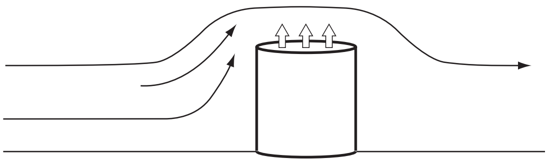

Figure 2.7 The effect of wind turbulence on a raised rain gauge. An area of reduced pressure (and uplift) develops above the gauge in a similar manner to an aircraft wing. This reduces the rain gauge catch.

both the wind speed and the raindrop diameter (Nešpor and Sevruk, 1999). At wind speeds of 20 km/hr (Beaufort scale 2) the loss could be up to 20 per cent, and in severe winds of 90 km/hr (Beaufort scale 8) up to 40 per cent (Bruce and Clark, 1980; Rodda and Smith, 1986). The higher a gauge is from the surface the greater the loss of accuracy. This creates a major problem for gauges in areas that receive large snowfalls as they need to be raised to avoid surface coverage.

One method of addressing these turbulence difficulties is through the fitting of a shield to the rain gauge (see Figure 2.8). A rain gauge shield can take many forms but is often a series of batons surrounding the gauge at its top height. The shield acts as a calming measure for wind around the gauge and has been shown to greatly improve rain gauge accuracy.

The optimum rain gauge design

There is no perfect rain gauge. The design of the best gauge for a site will be influenced by the individual conditions at the site (e.g. prevalence of snowfall, exposure, etc.). A rain gauge with a non-splash surround, such as in Figure 2.6, can give very accurate measurement but it is prone to coverage by heavy snowfall so cannot always be used. The non-splash surround allows adjacent rainfall to pass through (negating splash) but acts as an extended soil surface for the wind, thereby eliminating the turbulence problem from raised gauges. This may be the closest that it is possible to get to measuring the amount of rainfall that would have fallen on a surface if the rain gauge were not there.

The standard UK Meteorological Office rain gauge has been adopted around the world (although not everywhere) as a compromise between the factors influencing rain gauge accuracy. It is a brass-rimmed rain gauge of 5 inches (127 mm) diameter standing 1 foot (305 mm) above the ground. The lack of height above ground level is a reflection of the low incidence of snowfall in the UK; in countries such as Russia and Canada, where winter snowfall is the norm, gauges may be raised as high as 2 m above the surface. There is general recognition that the UK standard rain gauge is not the best design for hydrology, but it does represent a reasonable compromise. There is a strong argument to be made against changing its design. Any change in the measurement instrument would make an analysis of past rainfall patterns difficult due to the differing accuracy.

Siting of a rain gauge

Once the best measurement device has been chosen for a location there is still a considerable measurement error that can occur through incorrect siting. The major problem of rain gauge siting in hydrology is that the scientist is trying to measure the rainfall at a location that is representative of a far greater area. It is extremely important that the measurement location is an appropriate surrogate for the larger area. If the area of interest is a forested catchment then it is reasonable to place your rain gauge beneath the forest canopy; likewise, within an urban environment it is reasonable to expect interference from buildings because this is what is happening over the larger area. What is extremely important is that there are enough rain gauges to try and quantify the spatial and temporal variations.

The rule-of-thumb method for siting a rain gauge is that the angle when drawn from the top of the rain gauge to the top of the obstacle is less than 30° (see Figure 2.9). This can be approximated as at least twice the height of the obstacle away from the gauge. Care needs to be taken to allow for the future growth of trees so that at all times during the rainfall record the distance apart is at least twice the height of an obstacle.

Gauges for the continuous measurement of rainfall

The standard UK Meteorological Office rain gauge collects water beneath its funnel and this volume is read once a day. Often in hydrology the data needs to be measured at a finer timescale than this, particularly in the case of individual storms which often last much less than a day. The most common modern method for measuring continuous rainfall uses a tipping-bucket rain gauge. These are very simple devices that can be installed relatively cheaply, although they do require a data-logging device nearby. The principle behind the tipping-bucket rain gauge is that as the rain falls it fills up a small 'bucket' that is attached to another 'bucket' on a balanced cross arm (see Figure 2.10). The 'buckets' are very small plastic containers at the end of each cross arm. When the bucket is full it tips the balance so that the full bucket is lowered down and empties out. At the time of tipping a magnet attached to the balance arm closes a small reed switch which sends an electrical signal to a data-logging device. This then records the exact time of the tipped bucket. If the rain continues to fall it fills the bucket on the other end of the cross

Figure 2.9 Siting of a rain gauge away from obstructions.

Figure 2.10 The insides of a tipping-bucket rain gauge. The 'buckets' are the small white, triangular reservoirs. These are balanced and when full they tip over bringing the black arm past the other stationary arm. In doing so a small electrical current is passed to a data logger.

arm until it too tips the balance arm, sending another electrical impulse to the data logger. In this way a near continuous measurement of rainfall with time can be obtained.

It is important that the correct size of tipping bucket is used for the prevailing conditions. If the buckets are too small then a very heavy rainfall event will cause them to fill too quickly and water be lost through overspill while the mechanism tips. If the buckets are too large then a small rainfall event may not cause the cross arm to tip and the subsequent rainfall event will appear larger than it actually was. The tipping-bucket rain gauge shown in Figure 2.10 has an equivalent depth of 0.2 mm of rain which works well for field studies in south-east England.

Snowfall measurement

The measurement of snowfall has similar problems to those presented by rainfall, but they are often more extreme. There are two methods used for measuring snowfall: using a gauge like a rain gauge; or measuring the depth that is present on the ground. Both of these methods have very large errors associated with them, predominantly caused by the way that snow falls through the atmosphere and is deposited on the gauge or ground. Most, although not all, snowflakes are more easily transported by the wind than raindrops. When the snow reaches the ground it is easily blown around in a secondary manner (drifting). This can be contrasted to liquid water where, upon reaching the ground, it is either absorbed by the soil or moves across the surface. Rainfall is very rarely picked up by the wind again and redistributed in the manner that drifting snow is. For the snow gauge this presents problems that are analogous to rain splash. For the depth gauge the problem is due to uneven distribution of the snow surface: it is likely to be deeper in certain situations.

Rain gauge modification to include snowfall

One modification that needs to be made to a standard rain gauge in order to collect snowfall is a heated rim so that any snow falling on the gauge melts to be collected as liquid water. Failure to have a heated rim may mean that the snow builds up on the gauge surface until it overflows. Providing a heated rim is no simple logistic exercise as it necessitates a power source (difficult in remote areas) and the removal of collected water well away from the heat source to minimise evaporation losses.

A second modification is to raise the gauge well above ground level so that as snow builds up the gauge is still above this surface. Unfortunately the raising of the gauge leads to an increase in the turbulence errors described for rain gauges. For this reason it is normal to have wind deflectors or shields surrounding the gauge.

Snow depth

The simplest method of measuring snow depth is the use of a core sampler. This takes a core of snow, recording its depth at the same time, that can then be melted to derive the water equivalent depth. It is this that is of importance to a hydrologist. The major difficulties of a core sample are that it is a non-continuous reading (similar to daily rainfall measurement), and the position of coring may be critical (because of snow drifting).

A second method of measuring snow depth is to use a **snow pillow**. This is a method for measuring snow accumulation, a form of water storage, hence it is described in Chapter 4 (p. 75).

Forest rainfall measurement

The most common method of assessing the amount of canopy interception loss is to measure the precipitation above and below a canopy and assume that the difference is from interception. Stated in this way it sounds a relatively simple task but in reality it is fraught with difficulty and error. Durocher (1990) provides a good example of the instrumentation necessary to measure canopy interception, in this case for a deciduous woodland plot.

Above-canopy precipitation

To measure above-canopy precipitation a rain gauge may be placed on a tower above the canopy. The usual rain gauge errors apply here, but especially the exposure to the wind. As described in Chapter 3, the top of a forest canopy tends to be rough and is very good for allowing turbulent transfer of evaporated water. The turbulent air is not so good for measuring rainfall! An additional problem for any long-term study is that the canopy is not static; the tower needs to be raised every year so that it remains above the growing canopy.

One way around the tower problem is to place a rain gauge in a nearby clearing and assume that what falls there is the same amount as directly above the canopy nearby. This is often perfectly reasonable to assume, particularly for long-term totals, but care must be taken to ensure the clearing is large enough to avoid obstruction from nearby trees (see Figure 2.9).

Throughfall

Throughfall is the hardest part of the forest hydrological cycle to measure. This is because a forest canopy is normally variable in density and therefore throughfall is spatially heterogeneous. One common method is to place numerous rain gauges on the forest floor in a random manner. If you are interested in a long-term study then it is reasonable to keep the throughfall gauges in fixed positions. However, if the study is investigating individual storm events then it is considered best practice to move the gauges to new random positions between storm events. In this way the throughfall catch should not be influenced by gauge position. To derive an average throughfall figure it is necessary to come up with a spatial average in the same manner as for areal rainfall estimates (see below).

To overcome the difficulty of a small sampling area (rain gauge) measuring something notoriously variable (throughfall), some investigators have used either troughs or plastic sheeting. Troughs collect over a greater area and have proved to be very effective (see Figure 2.11). Plastic sheeting is the ultimate way of collecting throughfall over a large area, but has several inherent difficulties. The first is purely logistical in that it is difficult to install and maintain, particularly to make sure there are no rips. The second is that by having an impervious layer above the ground there is no, or very little, water entering the soil. This might not be a problem for a short-term study but is over the longer term, especially if the investigator is interested in the total

Figure 2.11 Throughfall troughs sitting beneath a pine tree canopy. This collects rain falling through the canopy over the area of the trough. It is sloping so that water flows to a collection point.

water budget. It may also place the trees under stress through lack of water, thus leading to an altered canopy.

Stemflow

The normal method of measuring stemflow is to place collars around a tree trunk that capture all the water flowing down the trunk. On trees with smooth bark this may be relatively simple but is very difficult on rough bark such as found on many conifers. It is important that the collars are sealed to the tree so that no water can flow underneath and that they are large enough to hold all the water flowing down the trunk. The collars should be sloped to one side so that the water can be collected or measured in a tipping-bucket rain gauge. Maintenance of the collars is very important as they easily clog up or become appropriate resting places for forest fauna such as slugs!

MOVING FROM POINT MEASUREMENT TO SPATIALLY DISTRIBUTED ESTIMATION

The measurement techniques described here have all concentrated on measuring rainfall at a precise location (or at least over an extremely small area). In reality the hydrologist needs to know how much precipitation has fallen over a far larger area, usually a catchment. To move from point measurements to a spatially distributed estimation it is necessary to employ some form of spatial averaging. The spatial averaging must attempt to account for an uneven spread of rain gauges in the catchment and the various factors that we know influence rainfall distribution (e.g. altitude, aspect and slope). A simple arithmetic mean will only work where a catchment is sampled by uniformly spaced rain gauges and where there is no diversity in topography. If these conditions were ever truly met then it is unlikely that there would be more than one rain gauge sampling the area. Hence it is very rare to use a simple averaging technique.

There are different statistical techniques that address the spatial distribution issues, and with the growth in use of **Geographic Information Systems (GIS)** it is often a relatively trivial matter to do the calculation. As with any computational task it is important to have a good knowledge of how the technique works so that any shortcomings are fully understood. Three techniques are described here: **Thiessen's polygons**, the **hypsometric method** and the **isohyetal method**. These methods are explored further in a Case Study on p. 31.

Thiessen's polygons

Thiessen was an American engineer working around the start of the twentieth century who devised a simple method of overcoming an uneven distribution of rain gauges within a catchment (very much the norm). Essentially Thiessen's polygons attach a representative area to each rain gauge. The size of the representative area (a polygon) is based on how close each gauge is to the others surrounding it.

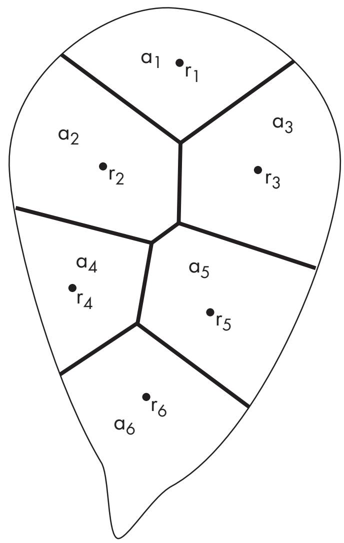

Figure 2.12 Thiessen's polygons for a series of rain gauges (r_i) within an imaginary catchment. The area of each polygon is denoted as a_i. Locations of rain gauges are indicated by bullet points.

Each polygon is drawn on a map; the boundaries of the polygons are equidistant from each gauge and drawn at a right angle (orthogonal) to an imaginary line between two gauges (see Figure 2.12). Once the polygons have been drawn the area of each polygon surrounding a rain gauge is found. The spatially averaged rainfall (R) is calculated using formula 2.1:

$$R = \sum_{i=1}^{n} \frac{r_i a_i}{A} \tag{2.1}$$

where r_i is the rainfall at gauge i, a_i is the area of the polygon surrounding rain gauge i, and A is the total catchment area.

The **areal rainfall** value using Thiessen's polygons is a weighted mean, with the weighting being based upon the size of each representative area (polygon). This technique is only truly valid where the topography is uniform within each polygon so that it can be safely assumed that the rainfall distribution is uniform within the polygon. This would suggest that it can only work where the rain gauges are located initially with this technique in mind (i.e. *a priori*).

Hypsometric method

Since it is well known that rainfall is positively influenced by altitude (i.e. the higher the altitude the greater the rainfall) it is reasonable to assume that knowledge of the catchment elevation can be brought to bear on the spatially distributed rainfall estimation problem. The simplest indicator of the catchment elevation is the hypsometric (or hypsographic) curve. This is a graph showing the proportion of a catchment above or below a certain elevation. The values for the curve can be derived from maps using a planimeter or using a digital elevation model (DEM) in a GIS.

The hypsometric method of calculating spatially distributed rainfall then calculates a weighted average based on the proportion of the catchment between two elevations and the measured rainfall between those elevations (equation 2.2).

$$R = \sum_{j=1}^{m} r_j p_j \tag{2.2}$$

where r_j is the average rainfall between two contour intervals and p_j is the proportion of the total catchment area between those contours (derived from the hypsometric curve). The r_j value may be an average of several rain gauges where there is more than one at a certain contour interval. This is illustrated in Figure 2.13 where the shaded area (a_3) has two gauges within it. In this case the r_j value will be an average of r_4 and r_5.

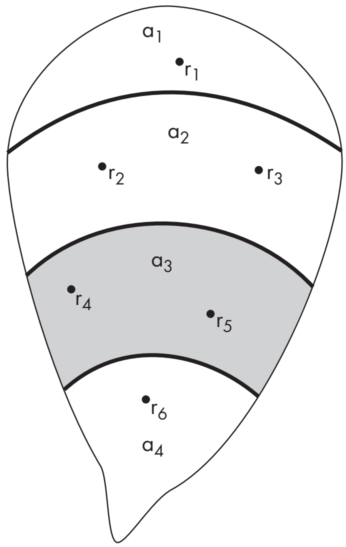

Figure 2.13 Calculation of areal rainfall using the hypsometric method. The shaded region is between two contours. In this case the rainfall is an average between the two gauges within the shaded area. Locations of rain gauges are indicated by bullet points.

Intuitively this is producing representative areas for one or more gauges based on contours and spacing, rather than just on the latter as for Thiessen's polygons. There is an inherent assumption that elevation is the only topographic parameter affecting rainfall distribution (i.e. slope and aspect are ignored). It also assumes that the relationship between altitude and rainfall is linear, which is not always the case and warrants exploration before using this technique.

Isohyetal and other smoothed surface techniques

Where there is a large number of gauges within a catchment the most obvious weighting to apply on a mean is based on measured rainfall distribution rather than on surrogate measures as described above. In this case a map of the catchment rainfall distribution can be drawn by interpolating between the rainfall values, creating a smoothed rainfall surface. The traditional isohyetal method involved drawing isohyets (lines of equal rainfall) on the map and calculating the area between each isohyet. The spatial average could then be calculated by equation 2.3

$$R = \sum_{i=1}^{n} \frac{r_i a_i}{A} \qquad (2.3)$$

where a_i is the area between each isohyet and r_i is the average rainfall between the isohyets. This technique is analogous to Figure 2.13, except in this case the contours will be of rainfall rather than elevation.

With the advent of GIS the interpolating and drawing of isohyets can be done relatively easily, although there are several different ways of carrying out the interpolation. The interpolation subdivides the catchment into small grid cells and then assigns a rainfall value for each grid cell (this is the smoothed rainfall surface). The simplest method of interpolation is to use a nearest neighbour analysis, where the assigned rainfall value for a grid square is proportional to the nearest rain gauges. A more complicated technique is to use **kriging**, where the interpolated value for each cell is derived with knowledge on how closely related the nearby gauges are to each other in terms of their co-variance. A fuller explanation of these techniques is provided by Bailey and Gatrell (1995).

An additional piece of information that can be gained from interpolated rainfall surfaces is the likely rainfall at a particular point within the catchment. This may be more useful information than total rainfall over an area, particularly when needed for numerical simulation of hydrological processes.

The difficulty in moving from the point measurement to a spatially distributed average is a prime example of the problem of scale that besets hydrology. The scale of measurement (i.e. the rain gauge surface area) is far smaller than the catchment area that is frequently our concern. Is it feasible to simply scale up our measurement from point sources to the overall catchment? Or should there be some form of scaling factor to acknowledge the large discrepancy? There is no easy answer to these questions and they are the type of problem that research in hydrology will be investigating in the twenty-first century.

RAINFALL INTENSITY AND STORM DURATION

Water depth is not the only rainfall measure of interest in hydrology; also of importance is the **rainfall intensity** and **storm duration**. These are simple to obtain from an analysis of rainfall records using frequency analysis. The rainfall needs to be recorded at a short time interval (i.e. an hour or less) to provide meaningful data.

Figure 2.15 shows the rainfall intensity for a rain gauge at Bradwell-on-Sea, Essex, UK. It is evident from the diagram that the majority of rain falls at very low intensity: 0.4 mm per hour is considered as light rain. This may be misleading as the rain gauge recorded rainfall every hour and the small amount of rain may have fallen during a shorter period than an hour i.e. a higher intensity but lasting for less than an hour. During the period of measurement there were recorded rainfall intensities greater than 4.4 mm/hr (maximum 6.8 mm/hr) but they were so few as to not show up on the histogram scale used in Figure 2.15. This may be a reflection of only two years of records being analysed, which

Case study

RAINFALL DISTRIBUTION IN A SMALL STUDY CATCHMENT

It is well known that large variations in rainfall occur over quite a small spatial scale. Despite this, there are not many studies that have looked at this problem in detail. One study that has investigated spatial variability in rainfall was carried out in the Plynlimon research catchments in mid-Wales (Clarke *et al*., 1973). In setting up a hydrological monitoring network in the Wye and Severn catchments thirty-eight rain gauges were installed to try and characterise the rainfall variation. The rainfall network had eighteen rain gauges in the Severn catchment (total area 8.7 km^2) and twenty gauges in the Wye (10.55 km^2).

The monthly data for a period between April 1971 and March 1973 were analysed to calculate areal average rainfall using contrasting methods. The results from this can be seen in Figure 2.14. The most startling feature of Figure 2.14 is the lack of difference in calculated values and that they follow no regular pattern. At times the arithmetic mean is greater than the others while in other months it is less. When the total rainfall for the two-year period is looked at, the Thiessen's calculation is 0.3 per cent less than the arithmetic mean, while the isohyetal method is 0.4 per cent less.

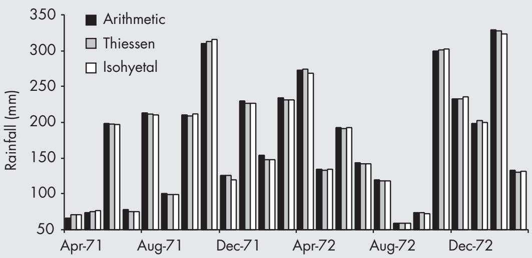

Figure 2.14 Areal mean rainfall (monthly) for the Wye catchment, calculated using three different methods.

Source: Data from Clarke *et al*. (1973)

When the data were analysed to see how many rain gauges would be required to characterise the rainfall distribution fully it was found that the number varied with the time period of rainfall and the season being measured. When monthly data were looked at there was more variability in winter rainfall than summer. For both winter and summer it showed that anything less than five rain gauges (for the Wye) increased the variance markedly.

A more detailed statistical analysis of hourly mean rainfall showed a far greater number of gauges were required. Four gauges would give an accuracy in areal estimate of around 50 per cent, while a 90 per cent accuracy would require 100 gauges (Clarke *et al*., 1973: 62).

The conclusions that can be drawn from the study of Clarke *et al*. (1973) are of great concern to hydrology. It would appear that even for a small catchment a large number of rain gauges are required to try and estimate rainfall values properly. This confirms the statement made at the start of this chapter: although rainfall is relatively straightforward to measure it is notoriously difficult to measure accurately and, to compound the problem, is also extremely variable within a catchment area.

introduces an extremely important concept in hydrology: the **frequency–magnitude** relationship. With rainfall (and runoff – see Chapters 5 and 6) the larger the rainfall event the less frequent we would expect it to be. This is not a linear relationship; as illustrated in Figure 2.15 the curve declines in a non-linear fashion. If we think of the relative frequency as a probability then we can say

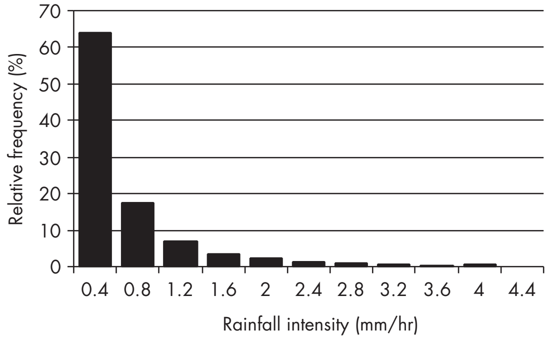

Figure 2.15 Rainfall intensity curve for Bradwell-on-Sea, Essex, UK. Data are hourly recorded rainfall from April 1995 to April 1997.

that the chances of having a low rainfall event are very high: a low magnitude–high frequency event. Conversely the chances of having a rainfall intensity greater than 5 mm/hr are very low (but not impossible): a high magnitude–low frequency event.

In Figure 2.16 the storm duration records for two different sites are compared. The Bradwell-on-Sea site has the majority of its rain events lasting one hour or less. In contrast the Ahoskie site has only 20 per cent of its storms lasting one hour or less but many more than Bradwell-on-Sea that last four hours or more. When the UK site rainfall and intensity curves are looked at together (i.e. Figures 2.15 and 2.16) it can be stated that Bradwell-on-Sea experiences a predominance of low intensity, short duration rainfall events and very few long duration, high intensity storms. This type of information is extremely useful to a hydrologist investigating the likely runoff response that might be expected for the rainfall regime.

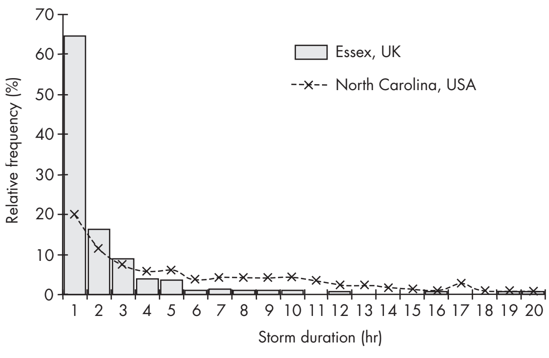

Figure 2.16 Storm duration curves. The bars are for the same data set as Figure 2.12 and the broken line for Ahoskie, North Carolina.

Source: Ahoskie data are redrawn from Wanielista (1990)

SURROGATE MEASURES FOR ESTIMATING RAINFALL

The difficulties in calculation of a spatially distributed precipitation value from point measurements make the direct estimation of areal precipitation an attractive proposition. There are two techniques that make some claim to achieving this: radar, and **satellite remote sensing**. These approaches have many similarities, but they differ fundamentally in the direction of measurement. Radar looks from the earth up into the atmosphere and tries to estimate the amount of precipitation falling over an area. Satellite remote sensing looks from space down towards the earth surface and attempts to estimate the amount of precipitation falling over an area.

Radar

The main use of ground-based radar is in weather forecasting where it is used to track the movement of rain clouds and fronts across the earth's surface. This in itself is interesting but does not provide the hydrological requirement of estimating how much rain is falling over an area.

There are several techniques used for radar, although they are all based on similar principles. Radar is an acronym (*r*adio *d*etection *a*nd *r*anging). A wave of electromagnetic energy is emitted from a unit on the ground and the amount of wave reflection and return time is recorded. The more water there is in a cloud the more electromagnetic energy is reflected back to the ground and detected by the radar unit. The quicker the reflected wave reaches back to ground the closer the cloud is to the surface. The most difficult part of this technique is in finding the best wavelength of electromagnetic radiation to emit and detect. It is important that the electromagnetic wave is reflected by liquid water in the cloud, but not atmospheric gases and/or changing densities of the atmosphere. A considerable amount of research effort has gone into trying to find the best wavelengths for ground-based radar to use. The solution appears to be that it is somewhere in the microwave band (commonly c-band), but that the exact wavelength depends on the individual situation being studied (Cluckie and Collier, 1991).

Studies have shown a good correlation between reflected electromagnetic waves and rainfall intensity. Therefore, this can be thought of as a surrogate measure for estimating rainfall. If an accurate estimate of rainfall intensity is required then a relationship has to be derived using several calibrating rain gauges. Herein lies a major problem: with this type of technique: there is no universal relationship that can be used to derive rainfall intensity from cloud reflectivity. An individual calibration has to be derived for each site and this may involve several years of measuring point rainfall coincidentally with cloud reflectivity. This is not a cheap option and the cost prohibits its widespread usage, particularly in areas with poor rain gauge coverage.

In Britain the UK Meteorological Office operates a series of fifteen weather radar with a 5 km resolution that provide images every 15 minutes. This is a more intensive coverage than could be expected in most countries. Although portable radar can be used for rainfall estimation, their usage has been limited by the high cost of purchase.

Satellite remote sensing

The atmosphere-down approach of satellite remote sensing is quite different from the ground-up approach of radar – fundamentally because the sensor is looking at the top of a cloud rather than the bottom. It is well established that a cloud most likely to produce rain has an extremely bright and cold top. These are the characteristics that can be observed from space by a satellite sensor. The most common form of satellite sensor is passive (this means it receives radiation from another source, normally the sun, rather than emitting any itself the way radar does) and detects radiation in the visible and infrared wavebands. **LANDSAT**, **SPOT** and **AVHRR** are examples of satellite platforms of this type. By sensing in the visible and infrared part of the electromagnetic spectrum the cloud brightness (visible) and temperature (thermal infrared) can be detected. This so-called 'brightness temperature' can then be related to rainfall intensity via calibration with point rainfall measurements, in a similar fashion to ground-based radar. One of the problems with this approach is that it is sometimes difficult to distinguish between snow reflecting light on the ground and clouds reflecting light in the atmosphere. They have similar brightness temperature values but need to be differentiated so that accurate rainfall assessment can be made.

Another form of satellite sensor that can be used is passive microwave. The earth emits microwaves (at a low level) that can be detected from space. When there is liquid water between the earth's surface and the satellite sensor (i.e. a cloud in the atmosphere) some of the microwaves are absorbed by the water. A satellite sensor can therefore detect the presence of clouds (or other bodies of water on the surface) as a lack of microwaves reaching the sensor. An example of a study using a satellite platform that can detect passive microwaves (SSM/I) is in Todd and Bailey (1995), who used the method to assess rainfall over the United Kingdom. Although there was some success in the method it is at a scale of little use to catchment scale hydrology as the best resolution available is around 10 × 10 km grid sizes.

Satellite remote sensing provides an indirect estimate of precipitation over an area but is still a long way from operational use. Studies have shown that it is an effective tool where there is poor rain gauge coverage (e.g. Kidd *et al.*, 1998), but in countries with high rain gauge density it does not improve estimation of areal precipitation. What is encouraging about the technique is that nearly all the world is covered by satellite imagery so that it can be used in sparsely gauged areas. The new generation of satellite platforms being launched in the early twenty-first century will have multiple sensors on them so it is feasible that they will be measuring visible, infrared and microwave wavebands simultaneously. This will improve the accuracy considerably but it must be borne in mind that it is an indirect measure of precipitation and will still require calibration to a rain gauge set.

PRECIPITATION IN THE CONTEXT OF WATER QUANTITY AND QUALITY

Precipitation, as the principal input into a catchment water balance, has a major part to play in water quantity and quality. By and large it is the spatial and temporal distribution of precipitation that drives the spatial and temporal distribution of available water. Rainfall intensity frequently controls the amount of runoff during a storm event (see Chapter 5) and the distribution of rain through the year controls the need for irrigation in an agricultural sytem.

The exception to this is in large river basins where the immediate rainfall distribution may have little bearing on the water flowing down the adjacent river. A good example is the Colorado River which flows through areas of extremely low rainfall in Utah and Arizona. The lack of rainfall in these areas has little bearing on the quantity of water in the Colorado River, it is governed by the precipitation (both rain and snow) falling in the Rocky Mountains well to the north-east.

The influence of precipitation on water quantity directly affects water quality through dilution. Where water quantity is high there is more water available to dilute any contaminants entering a river or groundwater system. It does not follow that high water quantity equates with high water quality but it has the potential to do so.

Precipitation also has a direct influence on water quality through scavenging of airborne pollutants which are then dissolved by the rain. The complex nature of a forest topography means that trees act as recipient surfaces for airborne pollutants. As rain falls onto the tree, salts that have formed on leaves and branches may be dissolved by the water, making the stemflow and throughfall pollutant-rich. This has been observed in field studies, particularly near the edge of tree stands (Neal *et al*., 1991).

The best known example of pollutant scavenging is **acid rain**. This is where precipitation in areas polluted by industrial smokestacks dissolves gases and absorbs particles that lower the acidity of the rain. Naturally rain is slightly acidic with pH between 5 and 6, due to the dissolving of carbon dioxide to form a weak carbonic acid. The burning of fossil fuels adds nitrogen oxides and sulphur oxides to the atmosphere, both of which are easily dissolved to form weak nitric and sulphuric acids. The burning of coal is particularly bad through the amount of sulphur dioxide produced, but any combustion will produce nitrogen oxides by the combination of nitrogen and oxygen (both already in the atmosphere) at high temperatures. In areas of the Eastern United States and Scandinavia rainfall has been recorded with pH values as low as 3 (similar to vinegar). In some situations this makes very little difference to overall water quality as the soil may have enough acid-buffering capability to absorb the acid rain. This is particularly true for limestone areas where the soil is naturally alkaline. However many soils do not have this buffering capacity due to their underlying geology (e.g. granite areas in the north-east of North America). In this situation the streams become acidic and this has an extremely detrimental effect on the aquatic fauna. The major reason for the impact on fish life is the dissolved aluminium that the acidic water carries; this interferes with the operation of gills and the fish effectively drown.

It is worth noting that the dissolving of nitrous oxide can have a positive benefit to plant life through the addition to the soils of nitrate which promotes plant growth (see Chapter 7). To give some scale to the impact of large atmospheric nitrogen inputs, Löye-Pilot *et al*. (1990) estimate that atmospheric nitrogen input into the Mediterranean Sea is of the same order of magnitude as the riverine input. It is estimated that 25 per cent of nitrogen inputs to the Baltic Sea come from the atmosphere (BSEP, 2005).

SUMMARY

Precipitation is the main input of water within a catchment water balance. Its measurement is fraught with difficulties and any small errors will be magnified enormously at the catchment scale. It is also highly variable in time and space. Despite these difficulties precipitation is one of the most regularly measured hydrological variables, and good rainfall records are available for many regions in the world. A forest canopy partitions rainfall into components that move at different rates towards the soil surface. The nature of the canopy (leaf size distribution and leaf area index) determines the impact that a canopy has on the water balance equation.

Analysis of rainfall can be carried out with respect to trying to find a spatial average or looking at the intensity and duration of storm events. Although there are techniques available for estimating precipitation their accuracy is not such that it is superior to a good network of precipitation gauges.

ESSAY QUESTIONS

1 **Describe the different factors affecting the spatial distribution of precipitation at differing scales.**

2 **How are errors in the measurement of rainfall and snowfall minimised?**

3 **Compare and contrast different techniques for obtaining a spatially averaged precipitation value (including surrogate measures).**

4 **Why is scale such an important issue in the analysis of precipitation in hydrology?**

5 **Describe a field experiment (including equipment) required to measure the water balance beneath a forest canopy.**

6 **Discuss the role of spatial scale in assessing the importance of a forest canopy within a watershed.**

FURTHER READING

Bailey, T.C. and Gatrell, A.C. (1995) *Interactive spatial data analysis*. Longman, Harlow.
Gives a modern view of spatial analysis, not necessarily just for precipitation.

Barry, R.G. and Chorley, R. (2003) *Atmosphere, weather and climate* (8th edition). Routledge, London.
Explains the principles of precipitation generation well.

Cheng, M. (2003) *Forest hydrology: an introduction to water and forests*. CRC Press, Boca Raton, Florida.
A more modern text than Lee (1980) which gives a good overview of forest hydrology.

Lee, R. (1980) *Forest hydrology*. Columbia University Press, New York.
A classic text giving an excellent overview of forest hydrological processes.

Strangeways, I. (2006) *Precipitation: theory, measurement and disribution*. Cambridge University Press.
A modern text on precipitation processes and measurement.

3

EVAPORATION

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the process of evaporation and what controls its rate.
- A knowledge of the techniques for measuring evaporation directly.
- A knowledge of the techniques used to estimate evaporation.

Evaporation is the transferral of liquid water into a gaseous state and its diffusion into the atmosphere. In order for this to occur there must be liquid water present and available energy from the sun or atmosphere. The importance of evaporation within the hydrological cycle depends very much on the amount of water present and the available energy, two factors determined by a region's climate. During winter months in humid-temperate climates evaporation may be a minor component of the hydrological cycle as there is very little available energy to drive the evaporative process. This alters during summer when there is abundant available energy and evaporation has the potential to become a major part of the water balance. The potential may be limited by the availability of liquid water during the dry months. This can be seen in extremely hot, arid climates where there is often plenty of available energy to drive evaporation but very little water to be evaporated. As a consequence the actual amount of evaporation is small.

It is the presence or lack of water at the surface that provides the major semantic distinction in definitions of the evaporative process. **Open water evaporation** (often denoted as E_o) is the evaporation that occurs above a body of water such as a lake, stream or the oceans. Figure 1.6 shows that at the global scale this is the largest source of evaporation, in particular from the oceans. **Potential evaporation** (*PE*) is that which occurs over the land's surface, or would occur if the water supply were unrestricted. This occurs when a soil is wet and what evaporation is able to happen occurs without a lack of water supply. **Actual evaporation** (E_t) is that which actually occurs (i.e. if there is not much available water it will be less than potential). When conditions are very wet (e.g. during a rainfall event) E_t will equal *PE*, otherwise it will be less than *PE*. In hydrology we are most interested in E_o and E_t but normally require *PE* to calculate the E_t value.

All of these definitions have been concerned with 'evaporation over a surface'. In hydrology the surface is either water (river, lake, ponds, etc.) or the land. The evaporation above a land surface occurs in two ways – either as actual evaporation from the soil matrix or **transpiration** from plants. The combination of these two is often referred to as **evapotranspiration**, although the term *actual evaporation* is essentially the same (hence the *t* subscript in E_t). Transpiration from a plant occurs as part of photosynthesis and respiration. The rate of transpiration is controlled by the opening or closing of stomata in the leaf. Transpiration can be ascertained at the individual plant level by instruments measuring the flow of water up the trunk or stem of a plant. Different species of plants transpire at different rates but the fundamental controls are the available water in the soil, the plant's ability to transfer water from the soil to its leaves and the ability of the atmosphere to absorb the transpired water.

Evaporation is sometimes erroneously described as the only loss within the water balance equation. The water balance equation is a mathematical description of the hydrological cycle and by definition there are no losses and gains within this cycle. What is meant by 'loss' is that evaporation is lost from the earth's surface, where hydrologists are mostly concerned with the water being. To a meteorologist, concerned with the atmosphere, evaporation can be seen as a gain. Evaporation although not a loss, can be viewed as the opposite of precipitation, particularly in the case of dewfall, a form of precipitation. In this case the **dewfall** (or negative evaporation) is a gain to the earth's surface.

EVAPORATION AS A PROCESS

It has already been said that evaporation requires an energy source and an available water supply to transform liquid water into water vapour. There is one more precondition: that the atmosphere be dry enough to receive any water vapour produced. These are the three fundamental parts to an understanding of the evaporation process. This was first understood by Dalton (1766–1844), an English physicist who linked wind speed and the dryness of the air to the evaporation rate.

Available energy

The main source of energy for evaporation is from the sun. This is not necessarily in the form of direct radiation, it is often absorbed by a surface and then re-radiated at a different wavelength. The normal term used to describe the amount of energy received at a surface is **net radiation** (Q^*), measured using a net radiometer. Net radiation is a sum of all the different heat fluxes found at a surface and can be described by equation 3.1.

$$Q^* = Q_S \pm Q_L \pm Q_G \tag{3.1}$$

where Q_S is the sensible heat flux; Q_L is the latent heat flux and Q_G is the soil heat flux.

Sensible heat is that which can be sensed by instruments. This is most easily understood as the heat we feel as warmth. The sensible heat flux is the rate of flow of that sensible heat.

Latent heat is the heat either absorbed or released during a phase change from ice to liquid water, or liquid water to water vapour. When water moves from liquid to gas this is a negative flux (i.e. energy is absorbed) whereas the opposite phase change (gas to liquid) produces a positive heat flux.

The **soil heat flux** is heat released from the soil having been previously stored within the soil. This is frequently ignored as it tends to zero over a 24-hour period and is a relatively minor contributor to net radiation.

Incoming solar radiation is filtered by the atmosphere so that not all the wavelengths of the electromagnetic spectrum are received at the earth's surface. Incoming radiation that reaches the surface is often referred to as short-wave radiation: visible light plus some bands of the infrared. This is not strictly true as clouds and water vapour in the atmosphere, plus trees and tall buildings above the surface, emit longer-wave radiation which also reaches the surface.

Outgoing radiation can be either reflected short-wave radiation or energy radiated back by the earth's surface. In the latter case this is normally in the infrared band and longer wavelengths and is referred to as long-wave radiation. This is a major source of energy for evaporation.

There are two other forms of available energy that under certain circumstances may be important sources in the evaporation process. The first is heat stored in buildings from an anthropogenic source (e.g. domestic heating). This energy source is often fuelled from organic sources and may be a significant addition to the heat budget in an urban environment, particularly in the winter months. The second additional source is **advective energy**. This is energy that originates from elsewhere (another region that may be hundreds or thousands of kilometres away) and has been transported to the evaporative surface (frequently in the form of latent heat) where it becomes available energy in the form of sensible heat. The best example of this is latent energy that arrives in cyclonic storm systems. In Chapter 1 it was explained that evaporating and condensing water is a major means of redistributing energy around the globe. The evaporation of water that contributes to cyclonic storms normally takes place over an ocean, whereas the condensation may occur a considerable distance away. At the time of evaporation, thermal energy (i.e. sensible heat) is transferred into latent energy that is then carried by the water vapour to the place of condensation where it is released as sensible heat once more. This 're-release' is often referred to as advective energy and may be a large energy source to drive further evaporation.

Water supply

Available water supply can be from water directly on the surface in a lake, river or pond. In this case it is open water evaporation (E_o). When the water is lying within soil the water supply becomes more complex. Soil water may evaporate directly, although it is normally only from the near surface. As the water is removed from the surface it sets up a soil moisture gradient that will draw water from deeper in the soil towards the surface, but it must overcome the force of gravity and the withholding force exerted by soil capillaries (see Chapter 4). In addition to this the water may be brought to the surface by plants using osmosis in their rooting system. The way that soil moisture controls the transformation from potential evaporation to actual evaporation is complex and will be discussed further later in this chapter.

The receiving atmosphere

Once the available water has been transformed into water vapour, using whatever energy source is available, it then must be absorbed into the atmosphere surrounding the surface. This process of *diffusion* requires that the atmosphere is not already saturated with water vapour and that there is enough buoyancy to move the water vapour away from the surface. These two elements can be assessed in terms of the **vapour pressure deficit** and atmospheric mixing.

Boyle's law tells us that the total amount of water vapour that may be held by a parcel of air is temperature and pressure dependent. The corollary of this is that for a certain temperature and air pressure it is possible to specify the maximum amount of water vapour that may be held by the parcel of air. We use this relationship to describe the **relative humidity** of the atmosphere (i.e. how close to fully saturated the atmosphere is). Another method of looking at the amount of water vapour in a parcel of air is to describe the *vapour pressure* and hence the *saturation vapour pressure*. The difference between the actual vapour pressure and the saturation vapour pressure is the *vapour pressure deficit* (vpd). The vpd is a measure of how much extra water vapour the atmosphere could hold assuming a constant temperature and pressure. The higher the vpd the more water can be absorbed from an evaporative surface.

Atmospheric mixing is a general term meaning how well a parcel of air is able to diffuse into the atmosphere surrounding it. The best indicator of atmospheric mixing is the wind speed at different heights above an evaporating surface. If the wind speed is zero the parcel of air will not move away from the evaporative surface and will 'fill' with water vapour. As the wind speed increases, the parcel of air will be moved quickly on to be replaced by another, possibly drier, parcel ready to absorb more water vapour. If the evaporative surface is large (e.g. a lake) it is important that the parcel of air moves up into the atmosphere, rather than directly along at the same level, so that there is drier air replacing it. This occurs through turbulent diffusion of the air. There is a greater turbulence associated with air passing over a rough surface than a smooth one, something that will be returned to in the discussion of evaporation estimation.

One way of thinking about evaporation is in terms of a washing line. The best conditions for drying your washing outside are on a warm, dry and windy day. Under these circumstances the evaporation from your washing (the available water) is high due to the available energy being high (it is a warm day), and the receiving atmosphere mixes well (it is windy) and is able to absorb much water vapour (the air is dry). On a warm and still day, or a warm and humid day washing does not dry as well (i.e. the evaporation rate is low). Understanding evaporation in these terms allows us to think about what the evaporation rate might be for particular atmospheric conditions.

Evaporation above a vegetation canopy

Where there is a vegetation canopy the evaporation above this surface will be a mixture of transpiration, evaporation from the soil and evaporation from wet leaves *(canopy interception or interception loss or wet leaf evaporation)*. The relative importance of these three evaporation sources will depend on the degree of vegetation cover and the climate at the site. In tropical rain forests transpiration is the dominant water loss but where there is a seasonal soil water deficit the influence of canopy interception loss becomes more important. This is illustrated by the data in Table 3.1 which contrasts the water balance for two Pinus radiata forests at different locations in New Zealand (with different climates).

Transpiration by a plant leads to evaporation from leaves through small holes (stomata) in the leaf. This is sometimes referred to as dry leaf evaporation. The influence of stomata on the transpiration rate is an interesting plant physiological phenomenon. Some

Table 3.1 Estimated evaporation losses from two *Pinus radiata* sites in New Zealand

	Puruki (Central North Island, NZ) (% annual rainfall in brackets)		*Balmoral (Central South Island, NZ) (% annual rainfall in brackets)*	
Annual rainfall	1,405 mm		870 mm	
Annual interception loss	370 mm	(26)	220 mm	(25)
Annual transpiration	705 mm	(50)	255 mm	(29)
Annual soil evaporation	95 mm	(7)	210 mm	(24)
Remainder (runoff + percolation)	235 mm	(17)	185 mm	(21)

Source: Data adapted from Kelliher and Jackson (2001) plants are very effective at shutting stomata when under water stress, and therefore limit their water usage. The water stress occurs when the vapour pressure deficit is high and there is a high evaporative demand. In this situation the stomata within a leaf can be likened to a straw. When you suck hard on a soft straw it creates a pressure differential between the inside and outside and the sides collapse in; therefore you cannot draw air easily through the straw. Stomata can act in a similar manner so that when the evaporative demand is high (sucking water vapour through the stomata) the stomata close down and the transpiration rate decreases. Some plant species shut their stomata when under evaporative stress (e.g. conifers) while others continue transpiring at high rates when the evaporative demand is high (e.g. many pasture species). The ability of plants to shut their stomata can influence the overall water budget as their overall evaporation is low. This is illustrated in the case study later in this chapter on using a lysimeter to measure tussock evaporation.

It is the role of interception loss (wet leaf evaporation) that makes afforested areas greater users of water than pasture land (see Case Study on p. 42). This is because the transpiration rates are similar between pasture and forest but the interception loss is far greater from a forested area. There are two influences on the amount of interception loss from a particular site: canopy structure and meteorology.

Canopy structural factors include the storage capacity, the drainage characteristics of the canopy and the aerodynamic roughness of the canopy. The morphology of leaf and bark on a tree are important factors in controlling how quickly water drains towards the soil. If leaves are pointed upwards then there tends to be a rapid drainage of water towards the stem. Sometimes this appears as an evolutionary strategy by a plant in order to harvest as much water as possible (e.g. rhubarb and gunnera plants). Large broadleaved plants, such as oak (*Quercus*), tend to hold water well on their leaves while needled plants can hold less per leaf (although they normally have more leaves). Seasonal changes make a large difference within deciduous forests, with far greater interception losses when the trees have leaves than without. Table 3.2 illustrates the influence of plant morphology through the variation in interception found in different forest types and ages. The largest influence that a canopy has in the evaporation process is through the aerodynamic roughness of the top of the canopy. This means that as air passes over the canopy it creates a turbulent flow that is very effective at moving evaporated water away from the surface. The reason that forests have such high interception losses is because they have a lot of intercepting surfaces *and* they have a high aerodynamic roughness leading to high rates of diffusion of the evaporated water away from the leaf (Figure 3.1).

Meteorological factors affecting the amount of interception loss are the rainfall characteristics. The rate at which rainfall occurs (intensity) and storm duration are critical in controlling the interception loss. The longer water stays on the canopy the greater the amount of interception loss. Also important will be the frequency of rainfall. Does

Table 3.2 Interception measurements in differing forest types and ages

Tree type	*Age*	*Interception (mm)*	*% of annual precipitation*
Deciduous hardwoods	100	254	12
Pinus strobus (White Pine)	10	305	15
Pinus strobus	35	381	19
Pinus strobus	60	533.4	26

Source: From Hewlett & Nutter (1969)

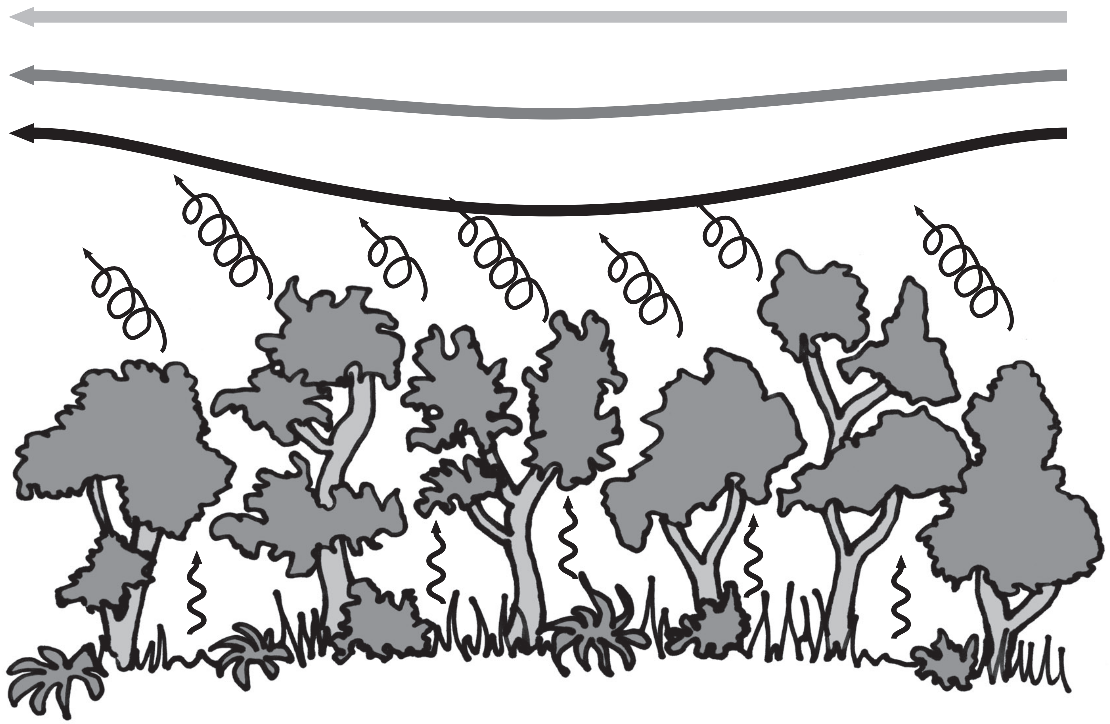

Figure 3.1 Factors influencing the high rates of interception loss from a forest canopy. The capacity of the leaves to intercept rainfall and the efficient mixing of water vapour with the drier air above leads to high evaporative losses (interception loss).

the canopy have time to dry out between rain events? If so, then the interception amount is likely to be higher. This is demonstrated in Figure 3.2 where the percentage of interception loss (interception ratio – broken line) is higher for small daily rainfall totals and the actual interception amount (solid line) reaches a maximum value of around 7mm even in the largest of daily rainfalls.

The amount of interception loss from an area is climate dependent. Calder (1990) used an amalgamation of different UK forest interception studies to show that there is a higher interception ratio (the interception loss divided by above-canopy rainfall) in drier than in wetter climates. The interception ratio ranges from 0.45 at 500 mm annual rainfall, to 0.27 at 2,700 mm annual rainfall. It is important to note that these interception ratio figures have considerable inter-annual variability.

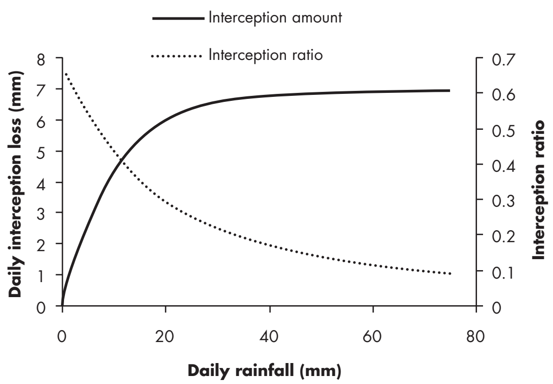

Figure 3.2 Empirical model of daily interception loss and the interception ratio for increasing daily rainfall. An interception ratio of 1.0 means all rainfall becomes interception loss.

Redrawn from Calder (1999)

Case study

FORESTS AND RAINFALL VS EVAPORATION

If you stand watching a forest during a warm summer shower it is common enough to see what appear to be clouds forming above the trees (see Plate 4). For many years it was believed that somehow trees attract rainfall and that cloud-forming was evidence of this phenomenon. As described by Pereira (1989) 'The worldwide evidence that hills and mountains usually have more rainfall and more natural forests than do adjacent lowlands has historically led to confusion of cause and effect'. This idea was taken further so that it became common practice to have forestry as a major land use in catchments that were being used to collect water for potable supply. In fact the cloud formation that is visible above a forest is a result of evaporation occurring from water sitting on the vegetation (intercepted rainfall). This 'wet leaf evaporation' can be perceived as a loss to the hydrologist as it does not reach the soil surface and contribute to possible streamflow. Throughout the latter half of the twentieth century there was considerable debate on how important wet leaf evaporation is.

One of the first pieces of field research to promote the idea of canopy interception being important was undertaken at Stocks Reservoir, Lancashire, UK. Law (1956) studied the water balance of an area covered with conifers (Sitka spruce) and compared this to a similar area covered with grassland. The water balance was evaluated for areas isolated by impermeable barriers with evaporation left as the residual (i.e. rainfall and runoff were measured and soil moisture assumed constant by looking at yearly values). Law found that the evaporation from the forested area was far greater than that for the pasture and he speculated that this was caused by wet leaf evaporation – in particular that the wet leaf evaporation was far greater from the forested area as there was a greater storage capacity for the intercepted water. Furthermore, Law went on to calculate the amount of water 'lost' to reservoirs through wet leaf vegetation and suggested a compensation payment from the forestry owners to water suppliers.

Conventional hydrological theory at the time suggested that wet leaf evaporation was not an important part of the hydrological cycle because it compensated for the reduction in transpiration that occurred at the same time (e.g. Leyton and Carlisle, 1959; Penman, 1963). In essence it was believed that the evapotranspiration rate stayed constant whether the canopy was wet or dry.

Following the work of Law, considerable research effort was directed towards discovering whether the wet leaf/dry leaf explanation was responsible for discrepancies in the water balance between grassland and forest catchments. Rutter (1967) and Stewart (1977) found that wet leaf evaporation in forests may be up to three or four times that from dry leaf. In contrast to this, other work has shown that on grassland, wet leaf evaporation is approximately equal to dry leaf (McMillan and Burgy, 1960; McIlroy and Angus, 1964). In addition, transpiration rates for pasture have been found to be similar to that of forested areas. When all this evidence is added up it confirms Law's work that forested areas 'lose' more rainfall through evaporation of intercepted water than grassland areas.

However there is still a question over whether the increased wet leaf evaporation may lead to a higher regional rainfall; a form of water recycling. Bands *et al*. (1987) write that: 'Forests are associated with high rainfall, cool slopes or moist areas. There is some evidence that, on a continental scale, forests may form part of a hydrological feedback loop with evaporation contributing to further rainfall'. Most researchers conclude that in general there is little, if any, evidence that forests can increase rainfall. However Calder (1999: 24, 26) concludes, 'Although the effects of forests on rainfall are likely to be relatively small, they cannot be totally dismissed from a water resources perspective . . . Further research is required to determine the magnitude of the effect, particularly at the regional scale.'

MEASUREMENT OF EVAPORATION

In the previous chapter there has been much emphasis on the difficulties of measuring precipitation due to its inherent variability. All these difficulties also apply to the measurement of evaporation, but they pale into insignificance when you consider that now we are dealing with measuring the rate at which a gas (water vapour) moves away from a surface. Concentrations of gases in the atmosphere are difficult to measure, and certainly there is no gauge that we can use to measure total amounts in the same way that we can for precipitation.

In each of the process chapters in this book there is an attempt to distinguish between measurement and estimation techniques. In the case of evaporation this distinction becomes extremely blurred. In reality almost all the techniques used to find an evaporation rate are estimates, but some are closer to true measurement than others. In this section each technique will include a sub-section on how close to 'true measurement' it is.

Direct micro-meteorological measurement

There are three main methods used to measure evaporation directly: the eddy fluctuation (or correlation), aerodynamic profile, and **Bowen ratio** methods. These are all micro-meteorological measurement techniques and details on them can be found elsewhere (e.g. Oke, 1987). An important point to remember about them all is that they are attempting to measure how much water is being evaporated above a surface, a very difficult task.

The eddy fluctuation method measures the water vapour above a surface in conjunction with a vertical wind speed and temperature profiles. These have to be measured at extremely short timescales (e.g. microseconds) to account for eddies in vertical wind motion. Consequently, extremely detailed micrometeorological instrumentation is required with all instruments having a rapid response time. In recent years this has become possible with hot wire **anemometers** and extremely fine thermistor heads for thermometers. One difficulty is that you are necessarily measuring over a very small surface area and it may be difficult to scale up to something of interest to catchment-scale hydrology.

The aerodynamic profile (or turbulent transfer) method is based on a detailed knowledge of the energy balance over a surface. The fundamental idea is that by calculating the amount of energy available for evaporation the actual evaporation rate can be determined. The measurements required are changes in temperature and humidity giving vertical humidity gradients. To use this method it must be assumed that the atmosphere is neutral and stable, two conditions that are not always applicable.

The Bowen ratio method is similar to the aerodynamic profile method but does not assume as much about the atmospheric conditions. The Bowen ratio is the ratio of sensible heat to latent heat and requires detailed measurement of net radiation, soil heat flux, temperature and humidity gradient above a surface. These measurements need to be averaged over a 30-minute period to allow the inherent assumptions to apply.

All of these micro-meteorological approaches to measuring evaporation use sophisticated instruments that are difficult to leave in the open for long periods of time. In addition to this they are restricted in their spatial scope (i.e. they only measure over a small area). With these difficulties it is not surprising that they tend to be used at the very small scale, mostly to calibrate estimation techniques (see pp. 46–52). They are accurate in the assessment of an evaporation rate, hence their use as a standard for the calibration of estimation techniques. The real problem for hydrology is that it is not a robust method that can be relied on for long periods of time.

Indirect measurement (water balance techniques)

Evaporation pans

The most common method for the measurement of evaporation is using an **evaporation pan** (see Figure 3.3). This is a large pan of water with a water depth measuring instrument or weighing device underneath that allows you to record how much water is lost through evaporation over a time period. This technique is actually a manipulation of the water balance equation, hence the terminology used here of a water balance technique. An evaporation pan is constructed from impervious material and the water level is maintained below the top so that no seepage or leakage occurs. This eliminates runoff (Q term) from the water balance. Therefore it can be assumed that any change in storage is related to either evaporative loss or precipitation gain. This means that the water balance equation can be rearranged as shown in equation 3.2.

$$E = \Delta S - P \tag{3.2}$$

If there is a precipitation gauge immediately adjacent to the evaporation pan then the P term can be accounted for, leaving only the change in storage (ΔS) to be measured as either a weight loss or a drop in water depth. At a standard meteorological station the evaporation is measured daily as the change in water depth. For a finer temporal resolution (e.g. hourly) there are load cell instruments available which measure and record the weight at regular intervals.

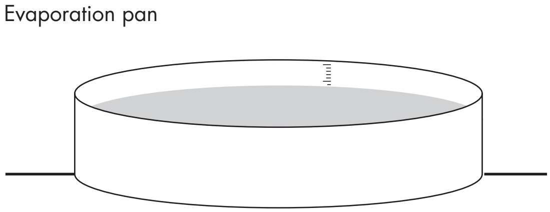

Figure 3.3 An evaporation pan. This sits above the surface (to lessen rain splash) and has either an instrument to record water depth or a continuous weighing device, to measure changes in volume.

An evaporation pan is filled with water, hence you are measuring E_o, the open water evaporation. Although this is useful, there are severe problems with using this value as an indicator of actual evaporation (E_t) in a catchment. The first problem is that E_o will normally be considerably higher than E_t because the majority of evaporation in a catchment will be occurring over a land surface where the available water is contained within soil and may be limited. This will lead to a large overestimation of the actual evaporation. This factor is well known and consequently evaporation pans are rarely used in catchment water balance studies, although they are useful for estimating water losses from lakes and reservoirs.

There are also problems with evaporation pans that make them problematic even for open water evaporation estimates. A standard evaporation pan, called a Class A evaporation pan, is 1,207 mm in diameter and 254 mm deep. The size of the pan makes them prone to the 'edge effect'. As warm air blows across a body of water it absorbs any water vapour evaporated from the surface. Numerous studies have shown that the evaporation rate is far higher near the edge of the water than towards the centre where the air is able to absorb less water vapour (this also applies to land surfaces). The small size of an evaporation pan means that the whole pan is effectively an 'edge' and will have a higher evaporation rate than a much larger body of water. A second, smaller, problem with evaporation pans is that the sides, and the water inside, will absorb radiation and warm up quicker than in a much larger lake, providing an extra energy source and greater evaporation rate.

To overcome the edge effect, empirical (i.e. derived from measurement) coefficients can be used which link the evaporation pan estimates to larger water body estimates. Doorenbos and Pruitt (1975) give estimates for these coefficients that require extra information on upwind fetch distance, wind run and relative humidity at the pan (Goudie *et al*., 1994). Grismer *et al*. (2002) provide empirical relationships linking pan evaporation measurements to potential evapotranspiration, i.e. from a vegetated surface not open water evaporation.

Lysimeters

A **lysimeter** takes the same approach to measurement as the evaporation pan, the fundamental difference being that a lysimeter is filled with soil and vegetation as opposed to water (see Figure 3.4). This difference is important, as E_t rather than E_o is being indirectly measured. A lysimeter can also be made to blend in with the surrounding land cover, lessening the edge effect described for an evaporation pan.

There are many versions of lysimeters in use, but all use some variation of the water balance equation to estimate what the evaporation loss has been. One major difference from an evaporation pan is that a lysimeter allows percolation through the bottom, although the amount is measured. Percolation is necessary so that the lysimeter mimics as closely as possible the soil surrounding it; without any it would fill up with water. In the same manner as an evaporation pan it is necessary to measure the precipitation input immediately adjacent to the lysimeter. Assuming that the only runoff (Q) is through percolation, the water balance equation for a lysimeter i shown in equation 3.3.

$$E = \Delta S - P - Q \qquad (3.3)$$

A lysimeter faces similar problems to a rain gauge in that it is attempting to measure the evaporation that would be lost from a surface if the lysimeter were not there. The difference from a rain gauge is that what is contained in the lysimeter should closely match the surrounding plants and soil. Although it is never possible to recreate the soil and plants within a lysimeter perfectly, a close approximation can be made and this represents the best efforts possible to measure evaporation. Although lysimeters potentially suffer from the same edge effect as evaporation pans, the ability to match the surrounding vegetation means there is much less of an edge effect.

A *weighing lysimeter* has a weighing device underneath that allows any change in storage to be monitored. This can be an extremely sophisticated device (e.g. Campbell and Murray, 1990; Yang *et al*., 2000), where percolation is measured continuously using the same mechanism for a tipping-bucket rain gauge, weight changes are recorded continuously using a hydraulic pressure gauge, and precipitation is measured simultaneously. A variation on this is to have a series of small weighing lysimeters (such as small buckets) that can be removed and weighed individually every day to provide a record of weight loss. At the same time as weighing, the amount of percolation needs to be recorded. This is a very cheap way of estimating evaporation loss for a study using low technology.

Without any instrument to weigh the lysimeter (this is sometimes referred to as a *percolation gauge*) it must be assumed that the change in soil moisture over a period is zero and therefore evaporation equals rainfall minus runoff. This may be a reasonable assumption over a long time period such as a year where the soil storage will be approximately the

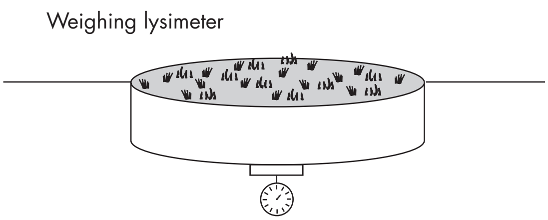

Figure 3.4 A weighing lysimeter sitting flush with the surface. The cylinder is filled with soil and vegetation similar to the surroundings.

same between two winters. An example of this type of lysimeter was the work of Law who investigated the effect that trees had on the water balance at Stocks Reservoir in Lancashire, UK (Law, 1956; see Case Study on p. 42).

A well-planned and executed lysimeter study probably provides the best information on evaporation that a hydrologist could find. However, it must be remembered that it is not evaporation that is being measured in a lysimeter – it is almost everything else in the water balance equation, with an assumption being made that whatever is left must be caused by evaporation. One result of this is that any errors in measurement of precipitation and/or percolation will transfer and possibly magnify into errors of evaporation measurement.

ESTIMATION OF EVAPORATION

The difficulties in measuring evaporation using either micro-meteorological instruments (problematic when used over long time periods and at the catchment scale) or water balance techniques (accumulated errors and small scale) has led to much effort being placed on estimating evaporation rather than trying to actually measure it. Some of the techniques outlined below are complicated and this sometimes leads hydrologists to believe that they are measuring, rather than estimating, evaporation. What they are actually doing is taking climatological variables that are known to influence evaporation and simulating evaporation rates from these: an estimation technique. The majority of research effort in this field has been to produce models to estimate evaporation; however, more recently, satellite remote sensing has provided another method of estimating the evaporation flux.

The techniques described here represent a range of sophistication and they are certainly not all universally applicable. Almost all of these are concerned with estimating the potential evaporation over a land surface. As with most estimation techniques the hydrologist is required to choose the best techniques for the study situation. In order to help in this decision the various advantages and shortcomings of each technique are discussed.

Thornthwaite

Thornthwaite derived an empirical model (i.e. derived from measurement not theoretical understanding) linking average air temperature to potential evaporation. This is an inherently sensible link in that we know air temperature is closely linked to both available energy and the ability of air to absorb water vapour.

The first part of the Thornthwaite estimation technique (Thornthwaite, 1944, 1954) derives a monthly heat index (i) for a region based on the average temperature t (°C) for a month (equation 3.4).

$$i = \left(\frac{t}{5}\right)^{1.514} \qquad (3.4)$$

These terms are then summed to provide an annual heat index I (equation 3.5).

$$I = \sum_{j=1}^{12} i \qquad (3.5)$$

Thornthwaite then derived an equation to provide evaporation estimates based on a series of observed evaporation measurements (equation 3.6).

$$PE = 16b\left(\frac{10t}{I}\right)^{a} \qquad (3.6)$$

The a and b terms in this equation can be derived in the following ways. Term b is a correction factor to account for unequal day length between months. Its value can be found by looking up tables based on the latitude of your study site. Term a is calibrated as a cubic function from the I term such as is shown in equation 3.7.

$$a = 6.7 \times 10^{-7} I^3 - 7.7 \times 10^{-5} I^2 + 0.018\, I + 0.49 \qquad (3.7)$$

Case study

A LYSIMETER USED TO MEASURE EVAPORATION FROM TUSSOCK

A narrow-leafed tussock grass (*Chionochloa rigida*, commonly called 'snow' or 'tall tussock') covers large areas of the South Island of New Zealand. A field study of a catchment dominated by snow tussock (Pearce *et al*., 1984) showed high levels of baseflow (i.e. high levels of streamflow between storm events). Mark *et al*. (1980) used a percolation gauge under a single tussock plant and estimating evaporation, showed that the water balance can show a surplus. They suggested that this may be due to the tussock intercepting fog droplets that are not recorded as rainfall in a standard rain gauge (see Plate 3). The nature of a tussock leaf (long and narrow with a sharp point), would seem to be conducive to fog interception in the same manner as conifers intercepting fog. Another interpretation of the Mark *et al*. (1980) study is that the estimation of evaporation was incorrect. An understanding of the mechanisms leading to high baseflow levels is important for a greater understanding of hydrological processes leading to streamflow.

In order to investigate this further a large lysimeter was set up in two different locations. The lysimeter was 2 m in diameter and contained nine mature snow tussock plants in an undisturbed monolith, weighing approximately 8,000 kg. Percolating runoff was measured with a tipping-bucket mechanism and the whole lysimeter was on a beam balance giving a sensitivity of 0.054 mm (Figure 3.5). The rainfall was measured immediately adjacent to the lysimeter. Campbell and Murray (1990) show that although there were times when fog interception appeared to occur (i.e. the catch in the lysimeter was greater than that in the nearby rain gauge) this only accounted for 1 per cent of the total precipitation. The detailed micro-meteorological measurements showed that the tussock stomatal or canopy resistance term was very high and that the plants had an ability to stop transpiring when the water stress became too high (see earlier discussion on plant physiological response to evaporative stress). The conclusion from the study was that snow tussocks are conservative in their use of water, which would appear to account for the high baseflow levels from tussock-covered catchments (Davie *et al*., 2006).

Figure 3.5 Large weighing lysimeter at Glendhu being installed. The weighing mechanism can be seen underneath.

(Photograph courtesy of Barry Fahey)

The Thornthwaite technique is extremely useful as potential evaporation can be derived from knowledge of average temperature (often readily available from nearby weather stations) and latitude. There are drawbacks to its usage however; most notably that it only provides estimates of monthly evaporation. For anything at a smaller time-scale it is necessary to use another technique such as

Penman's (see below). There are also problems with using Thornthwaite's model in areas of high potential evaporation. The empirical nature of the model means that it has been calibrated for a certain set of conditions and that it may not be applicable outside these. The Thornthwaite model has been shown to underestimate potential evaporation in arid and semi-arid regions (e.g. Acheampong, 1986). If the model is being applied in conditions different to Thornthwaite's original calibration (humid temperate regions) it is advisable to find out if any researcher has published different calibration curves for the climate in question.

Penman

Penman was a British physicist who derived a theoretical model of evaporation. Penman's first theoretical model was for open water evaporation and is shown in equations 3.8 and 3.9 (Penman, 1948):

$$E_o = \frac{\Delta Q^* + \gamma E_a}{\Delta + \gamma} \tag{3.8}$$

where an empirical relationship states that:

$$E_a = 2.6\delta_e\left(1 + \frac{u}{1.862}\right) \tag{3.9}$$

and Q^* = net radiation (in evaporation equivalent units of mm/day)
Δ = rate of increase of the saturation vapour deficit with temperature (kPa/°C see Figure 3.6)
δ_e = vapour pressure deficit of the air (kPa)
γ = psychrometric constant (≈ 0.063 kPa/°C)
u = wind speed at 2 m elevation (m/s)

In his original formula Penman estimated net radiation from empirical estimates of short- and long-wave radiation. The formula given here requires observations of temperature, wind speed, vapour pressure (which can be derived from relative humidity) and net radiation and gives the evaporation in units of mm per day. All of these can be obtained from meteorological measurement (see p. 49). It is normal to use daily averages for these variables, although Shuttleworth (1988) has suggested that it should not be used for time steps of less than ten days. There are several different ways of presenting this formula, which makes it difficult to interpret between texts. The main difference is in whether the evaporation is a flux or an absolute rate. In the equation above terms like 'net radiation' have been divided by the amount of energy required to evaporate 1 mm of water (density of water (ρ) multiplied by the latent heat of vaporisation (λ)) to turn them into water equivalents. This means the equation derives an absolute value for evaporation rather than a flux.

Penman continued his work to consider the evaporation occurring over a vegetated surface (Penman and Scholfield, 1951), while others refined the work (e.g. van Bavel, 1966). Part of this refinement was to include a term for aerodynamic resistance (r_a) to replace E_a (equation 3.9). **Aerodynamic resistance** is a term to account for the way in which the water evaporating off a surface mixes with a potentially drier atmosphere above it through turbulent mixing. The rougher the canopy surface the greater degree of turbulent mixing that will occur since air passing over the surface is buffeted around by protruding objects. As it is a resistance term, the higher the value, the greater the resistance to mixing; therefore a forest has a lower value of r_a than smoother pasture. Some values of aerodynamic resistance for different vegetation types are given in Table 3.3.

Substituting the new aerodynamic resistance term into the Penman equation, and presenting the results as a water flux (kg of water per m^2 of area), the evaporation estimation equation can be written as equation 3.10.

$$PE = \frac{Q^*\Delta + \dfrac{\rho \cdot c_\rho \cdot \delta_e}{r_a}}{\lambda(\Delta + \gamma)} \tag{3.10}$$

where

Q^* = net radiation (W/m^2)
Δ = rate of increase of the saturation vapour pressure with temperature (kPa/°C) (see Figure 3.6)
ρ = density of air (kg/m^3)
c_p = specific heat of air at constant pressure (≈ 1,005 J/kg)
δ_e = vapour pressure deficit of the air (kPa)
λ = latent heat of vaporisation of water (J/kg) (see Figure 3.4)
γ = psychrometric constant (≈0.063 kPa/°C)
r_a = aerodynamic resistance to transport of water vapour (s/m) given by equation 3.11.

$$r_a = \frac{\left(\ln\left(\frac{z-d}{z_0}\right)\right)^2}{\kappa^2 u} \tag{3.11}$$

and

κ = Von Karman constant (≈ 0.41)
u = wind speed above canopy (m/s)
z = height of anemometer (m)
d = **zero plane displacement** (the height within a canopy at which wind speed drops to zero, often estimated at two-thirds of the canopy height) (m)
z_o = roughness length (often estimated at one eighth of vegetation height) (m)

Table 3.3 Estimated values of aerodynamic and stomatal resistance for different vegetation types

Vegetation type	*Aerodynamic resistance (r_a) (s/m)*	*Canopy resistance (r_c) (s/m)*
Pasture	30	50
Forest	6.5	112
Scrub	6.5	160
Tussock	7.0	120

Source: from Andrew and Dymond (2007).
NB although the values of canopy resistance are presented as fixed they actually vary considerably throughout a day and season

Although this formula looks complicated it is actually rather simple. It is possible to split the equation into two separate parts that conform to the understanding of evaporation already discussed. The available energy term is predominantly assessed through the net radiation (Q^*) term. Other terms in the equation relate to the ability of the atmosphere to absorb the water vapour (Δ, ρ, c_p, δ_e, λ, γ, this is referred to as the sensible heat transfer function) and the rate at which diffusion will absorb the water vapour into the atmosphere (κ, u, z_o, etc.).

Figure 3.6 shows the relationship between the saturated vapour pressure and temperature. The slope of this curve (Δ) is required in the Penman equation and its derivatives. This can be estimated from equation 3.12 using average air temperature (T, °C):

$$\Delta = \frac{2053.058\exp^{\frac{17.27T}{T+237.3}}}{(T+237.3)^2} \tag{3.12}$$

When using the Penman equation there are only four variables requiring measurement: net radiation, wind speed above the canopy, atmospheric humidity and temperature, which when combined will provide vapour pressure deficit (see Figures 3.6–3.8).

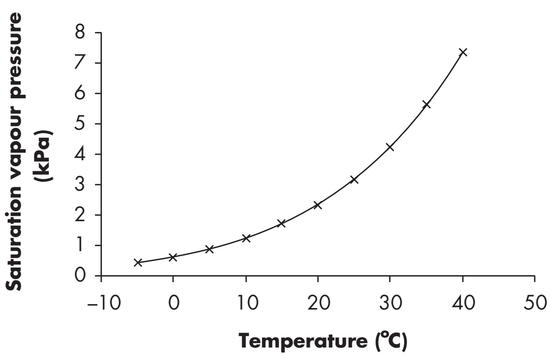

Figure 3.6 The relationship between temperature and saturation vapour pressure. This is needed to calculate the rate of increase of saturation vapour pressure with temperature (Δ). Equation 3.12 describes the form of this relationship.

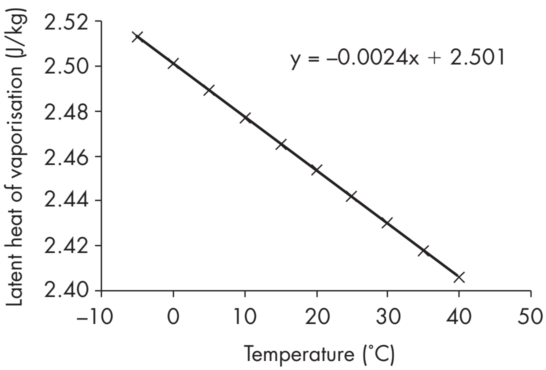

Figure 3.7 The relationship between temperature and latent heat of vaporisation.

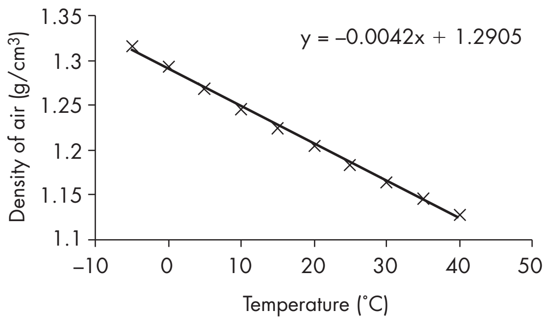

Figure 3.8 The relationship between air temperature and the density of air.

Every other term in the equation is either a constant, a simple relationship from another variable or can be measured once. Of these four variables net radiation is the hardest to obtain from meteorological stations as net radiometers are not common. There are methods of estimating net radiation from measurements of incoming solar radiation, surface **albedo** (or reflectivity) and day length (see Oke, 1987).

The modified Penman equation provides estimates of potential evaporation at a surface for time intervals much less than the monthly value from Thornthwaite. This makes it extremely useful to hydrology and it is probably the most widely used method for estimating potential evaporation values.

However, there are problems with the Penman equation which make it less than perfect as an estimation technique. The assumption is made that the soil heat flux is unimportant in the evaporation energy budget. This is often the case but is an acknowledged simplification that may lead to some overall error, especially when the time step is less than one day. It is normal practice to use Penman estimates at the daily time step; however, in some modelling studies they are used at hourly time steps.

One major problem with the Penman equation relates to its applicability in a range of situations and in particular in the role of advection, as discussed on p. 38. This is where there are other energy sources available for evaporation that cannot be assessed from net radiation. Calder (1990) shows the results from different studies in the UK uplands where evaporation rates vastly exceed the estimates provided by the Penman equation. The cause of this discrepancy is the extra energy provided by cyclonic storms coming onto Britain from the Atlantic Ocean, something that is poorly accounted for in the Penman equation. The part of the Penman equation dealing with the ability of the atmosphere to absorb the water vapour (sensible heat transfer function) does account for some advection but not if it is a major energy source driving evaporation and it is highly sensitive to the aerodynamic resistance term. This does not render the Penman approach invalid; rather, in applying it the user must be sure that net radiation is the main source of energy available for evaporation or the aerodynamic resistance term is well understood.

Simplifications to Penman

There have been several attempts made to simplify the Penman equation for widespread use. Slatyer and McIlroy (1961) separated out the evaporation caused by sensible heat and advection from that caused by radiative energy. Priestly and Taylor (1972) derived a simplified Penman formula for use in the large-scale estimation of evaporation, in the order of ‘several hundred kilometres’ where it can be argued that large-scale advection is not important. Their formula for potential evaporation is shown in equation 3.13.

$$PE = \alpha \frac{\left(Q^{*} - Q_G\right)\Delta}{\lambda\left(\Delta + \gamma\right)} \tag{3.13}$$

where Q_G is the soil heat flux term (often ignored by Penman but easily included if the measurements are available) and α is the Priestly–Taylor parameter, all other parameters being as defined earlier. The α term is an approximation of the sensible heat transfer function and was estimated by Priestly and Taylor (1972) to have a value of 1.26 for saturated land surfaces, oceans and lakes – that is to say, the sensible heat transfer accounts for 26 per cent of the evaporation over and above that from net radiation. This value of α has been shown to vary away from 1.26 (e.g. α = 1.21 in Clothier *et al*., 1982) but to generally hold true for large-scale areas without a water deficit.

Penman–Monteith

Monteith (1965) derived a further term for the Penman equation so that actual evaporation from a vegetated surface could be estimated. His work involved adding a canopy resistance term (r_c) into the Penman equation so that it takes the form of equation 3.14.

$$E_t = \frac{Q^{*}\Delta + \rho c_\rho \delta_e / r_a}{\lambda\left(\Delta + \gamma\left(1 + \frac{r_c}{r_a}\right)\right)} \tag{3.14}$$

Looking at the Penman–Monteith equation you can see that if r_c equals zero then it reverts to the Penman equation (i.e. actual evaporation equals potential evaporation). If the canopy resistance is high the actual evaporation rate drops to less than potential. Canopy resistance represents the ability of a vegetation canopy to control the rate of transpiration. This is achieved through the opening and closing of stomata within a leaf, hence r_c is sometimes referred to as stomatal resistance. Various researchers have established canopy resistance values for different vegetation types (e.g. Szeicz *et al*., 1969), although they are known to vary seasonally and in some cases diurnally. Rowntree (1991) suggests that for grassland under non-limiting moisture conditions the range of r_c should fall somewhere between 60 and 200 s/m. The large range is a reflection of canopy resistance being influenced by a plant's physiological response to variations in climatological conditions (see earlier discussion of stomatal control p. 40). Some values of canopy resistance for different vegetation types are given in Table 3.3.

Reference evaporation

The Penman–Monteith equation is probably the best evapotranspiration estimation method available. However for widespread use there is a need to have the stomatal resistance and aerodynamic resistance terms measured for a range of canopy covers at different stages of growth. To overcome this, the idea of reference evaporation has been introduced. This is the evaporation from a particular vegetation surface and the evaporation rate for another surface is related to this by means of crop coefficients. The Food and Agriculture Organisation (FAO) convened a group of experts who decided that the best surface for reference evaporation is close-cropped, well-watered grass. This is described in Allen *et al*. (1998) as a hypothetical reference crop with an assumed crop height of 0.12 m, a fixed canopy resistance of 70 s/m and an albedo of 0.23. Using these fixed values within the Penman–Monteith equation the reference evaporation (ET_o in mm/day) can be calculated from equation 3.15.

$$ET_o = \frac{0.408\Delta\left(Q * - Q_G\right) + \gamma \cdot \frac{900}{T + 273} u \cdot \delta_e}{\Delta + \gamma\left(1 + 0.34_u\right)} \tag{3.15}$$

where

- Q^* is net radiation at the crop surface (MJ/m²/day)

Q_G is soil heat flux density (MJ/m^2/day)
T is mean daily air temperature at 2 m height (°C)
u is wind speed at 2 m height (m/s)
δ_e is the saturation vapour pressure deficit (kPa)
Δ is the slope of the vapour pressure curve (kPa/°C)
γ is the psychrometric constant (kPa/°C)

The reference evapotranspiration, provides a standard to which evapotranspiration at different periods of the year or in other regions can be compared and evapotranspiration of other crops can be related (Allen *et al*., 1998). Scotter and Heng (2003) have investigated the sensitivity of the different inputs to the reference evaporation equation in order to show what accuracy of measurement is required.

Table 3.4 outlines some crop coefficients as set out by FAO (Allen *et al*., 1998). At the simplest level the evapotranspiration for a particular crop can be estimated by multiplying the crop coefficient with the reference evapotranspiration although there are more complex procedures outlined in Allen *et al*. (1998) which account for growth throughout a season and climatic variability. Where the crop coefficient values shown in Table 3.4 are higher than 1.0 it is likely that the aerodynamic roughness of the canopy makes for higher evaporation rates than for short grass. Where the values are less than 1.0 then the plants are exerting stomatal control on the transpiration rate.

Simple estimation of E_t from *PE* and soil moisture

Where there is no stomatal control exerted by plants (e.g. in a pasture) the relationship between actual evaporation (E_t) and potential evaporation (*PE*) is by and large driven by the availability of water. Over a land surface the availability of water can be estimated from the soil moisture content (see Chapter 4). At a simple level it is possible to estimate the relationship between potential and actual evaporation using soil moisture content as a measured variable (see Figure 3.9). In Figure 3.9 a value of 1 on the y-axis corresponds to actual precipitation equalling potential evaporation (i.e. available water is not a limiting factor on the evaporation rate). The exact position where this occurs will be dependent on the type of soil and plants on the land surface, hence the lack of units shown on the x-axis and the two different curves drawn. This type of simple relationship has been effective in determining actual evaporation rates in a model of soil water budgeting (e.g. Davie *et al*., 2001) but cannot be relied on for accurate modelling studies. It provides a very crude estimate of actual evaporation from knowledge of soil moisture and potential evaporation.

Table 3.4 Crop coefficients for calculating evapotranspiration from reference evapotranspiration

Crop type	*Crop coefficient (K_c)*	Comment
Beans and peas	1.05	Sometimes grown on stalks reaching 1.5 to 2 metres in height. In such cases, increased K_c values need to be taken.
Cotton	1.15–1.20	
Wheat	1.15	
Maize	1.15	
Sugar Cane	1.25	
Grapes	0.7	
Conifer forests	1.0	Confers exhibit substantial stomatal control. The K_c can easily reduce below the values presented, which represent well-watered conditions for large forests.
Coffee	0.95	

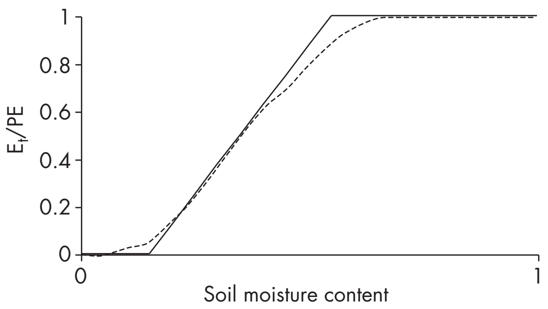

Figure 3.9 A hypothetical relationship between the measured soil moisture content and the ratio of actual evaporation to potential evaporation.

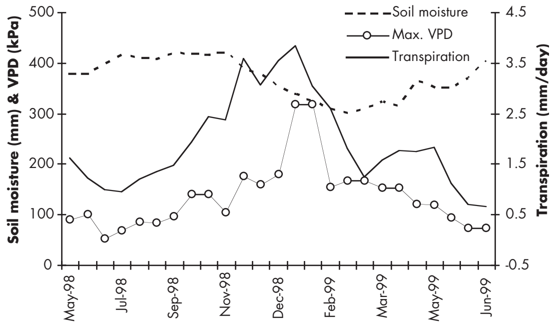

Figure 3.10 Time series of measured transpiration, measured soil moisture and estimated vapour pressure deficit for a forested site, near Nelson, New Zealand. NB as a Southern Hemisphere site the summer is from December until February.

Source: Data courtesy of Rick Jackson

The relationship between actual evaporation and soil moisture is not so simple where there is a vegetation type that exerts stomatal control on the evaporation rate (e.g. coniferous forest). In this case the amount of evaporation will be related to both soil moisture (available water) and the vapour pressure deficit (ability of the atmosphere to absorb water vapour). This is illustrated by Figure 3.10, a time series of soil moisture, transpiration and vapour pressure deficit for a stand of *Pinus radiata* in New Zealand. Transpiration was measured using sapflow meters on a range of trees; soil moisture was measured with a neutron probe and vapour pressure deficit was estimated from a nearby meteorological station. At the start of the summer period (Oct.–Nov. 1998) the soil moisture level is high and the transpiration rate climbs rapidly to a peak. Once it has reached the peak, the transpiration rate plateaus, despite the maximum vapour pressure deficit continuing to climb. During this plateau in transpiration rate the forest is exerting some stomatal control so that the transpiration doesn't increase by as much as the vapour pressure deficit. From January 1999 (the height of the Southern Hemisphere summer) the transpiration rate drops markedly. Initially this matches a drop in the maximum vapour pressure deficit but the transpiration rate continues to drop below early summer rates (with similar VPD values). This is the time that the lack of soil moisture is starting to limit the tree transpiration. Figure 3.10 illustrates the complex relationship between evaporation from a vegetated surface, the soil moisture conditions and the atmospheric conditions.

Remote sensing of evaporation

Water vapour is a greenhouse gas and therefore it interferes with radiation (i.e. absorbs and reradiates) from the earth's surface. Because of this the amount of water vapour in the atmosphere can be estimated using satellite remote sensing, particularly using passive microwave sensors. The difficulty with using this information for hydrology is that it is at a very large scale (often continental) and is concerned with the whole atmosphere not the near surface. In order to utilise satellites for estimation of evaporation a combined modelling and remote sensing approach is required. Burke *et al.* (1997) describe a combined Soil–Vegetation–Atmosphere–Transfer (SVAT) model that is driven by remotely sensed data. This type of approach can be used to estimate evaporation rates over a large spatial area relatively easily. Mauser and Schädlich (1998) provide a review of evaporation modelling at different scales using remotely sensed data.

Mass balance estimation

In the same manner that evaporation pans and lysimeters estimate evaporation rates, evaporation at the large scale (catchment or lake) can be estimated through the water balance equation. This is a relatively crude method, but it can be extremely effective over a large spatial and/or long temporal scale. The method requires accurate measurement of precipitation and runoff for a catchment or lake. In the case of a lake, change in storage can be estimated through lake-level recording and knowledge of the surface area. For a catchment it is often reasonable to assume that change in storage is negligible over a long time period (e.g. one year) and therefore the evaporation is precipitation minus runoff.

Canopy interception loss estimation

Empirical models that link rainfall to interception loss based on regression relationships of measured data sets have been developed for many different types of vegetation canopy (see Zinke (1967) and Massman (1983) for examples and reviews of these types of model). Some of these models used logarithmic or exponential terms in the equations but they all rely on having regression coefficients based on the vegetation type and climatic regime.

A more detailed modelling approach is the Rutter model (Rutter *et al*., 1971, 1975) which calculates an hourly water balance within a forest stand. The water balance is calculated, taking into account the rate of throughfall, stemflow, interception loss through evaporation and canopy storage. In order to use the model a detailed knowledge of the canopy characteristics is required. In particular the canopy storage and drainage rates from throughfall are required to be known; the best method for deriving these is through empirical measurement. The Rutter model treats the canopy as a single large leaf, although it has been adapted to provide a three-dimensional canopy (e.g. Davie and Durocher, 1997) that can then be altered to allow for changes and growth in the canopy.

At present, remote sensing techniques are not able to provide reasonable estimates of canopy interception. They do provide some useful information that can be incorporated into canopy interception models but cannot provide the detailed difference between above- and below-canopy rainfall. In particular, satellites can give good information on the type of vegetation and its degree of cover. Particular care needs to be taken over the term 'leaf area index' when reading remote sensing literature. Analysis of remotely sensed images can provide a good indication of the percentage vegetation cover for an area, but this is not necessarily the same as leaf area index – although it is sometimes referred to as such. Leaf area index is the surface area of leaf cover above a defined area divided by the surface area defined. As there are frequently layers of vegetation above the ground, the leaf area index frequently has a value higher than one. The percentage vegetation cover cannot exceed one (as a unitary percentage) as it does not consider the third dimension (height).

EVAPORATION IN THE CONTEXT OF WATER QUANTITY AND QUALITY

Evaporation, as the only loss away from the surface in the water balance equation, plays a large part in water quantity. The loss of water from soil through direct evaporation and transpiration has a direct impact on the amount of water reaching a stream during high rainfall (see Chapter 5) and also the amount of water able to infiltrate through into groundwater (see Chapter 4). The impact of evaporation on water quantity is not as great as for precipitation but it does have a significant part to play in the quantity and timing of water flowing down a river.

The influence of evaporation on water quality is mostly through the impurities left behind after water has evaporated. This may lead to a concentration of impurities in the water remaining behind (e.g. the Dead Sea between Israel and Jordan) or a build up of salts in soils (salination). This is discussed in more detail in Chapter 8.

SUMMARY

The evaporation process involves the transfer of water from a liquid state into a gaseous form in the atmosphere. For this to happen requires an available energy source, a water supply and the ability of the atmosphere to receive it. Evaporation is difficult to measure directly and there are various estimation techniques. These range from water budget techniques, such as evaporation pans and lysimeters, to modelling techniques, such as the Penman–Monteith equation. As a process, evaporation suffers from the same problems with measurement and estimation as does precipitation (i.e. extreme variability in space and time). This variability leads to difficulties in moving from point measurements to areal estimates such as are required for a catchment study. These can be overcome by using spatial averaging techniques or using evaporation estimations that assume a large base area (e.g. Priestly–Taylor). Forests have an important role to play in evaporation, particularly through interception loss. In general, more water is lost from a forested catchment than a non-forested catchment. This is through evaporation off wet leaves, but this is not always the case – there are cases where a tree canopy leads to more water in the catchment. The importance of canopy interception in a catchment water balance is dependent on the size and extent of vegetation cover found within a watershed.

ESSAY QUESTIONS

1 **Give a detailed account of the factors influencing evaporation rate above a forest canopy.**

2 **Compare and contrast the use of evaporation pans and lysimeters for measuring evaporation.**

3 **Outline the major evaporation estimation techniques and compare their effectiveness for your local environment.**

4 **Describe the factors that restrict actual evaporation (evapotranspiration) from equalling potential evaporation in a humid-temperate climate.**

FURTHER READING

Allen, R.G., Pereira L.S., Raes, D. and Smith, D. (1998) Crop evapotranspiration – Guidelines for computing crop water requirements. *FAO Irrigation and drainage paper 56* (available at www.fao.org/documents).

Brutsaert, W. (1982) *Evaporation into the atmosphere: theory, history, and applications*. Kluwer, Dordrecht.
A detailed overview of the evaporation process.

Calder, I.R. (1990) *Evaporation in the uplands*. J. Wiley & Sons, Chichester.
Although concerned primarily with upland evaporation it covers the issues of estimation well.

Cheng, M. (2003) Forest hydrology: an introduction to water and forests. CRC Press, Boca Raton, Florida.
An overview of forest hydrological processes, including evaporation and interception loss.

4

STORAGE

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the role of water stored below the ground (in both the saturated and unsaturated zones).
- An understanding of the role of snow and ice acting as a store for water.
- A knowledge of the techniques for measuring snow and ice and water beneath the ground.
- A knowledge of the techniques used to estimate the amount of water stored as soil moisture, groundwater and snow and ice.

The water balance equation, explained in Chapter 1, contains a storage term (S). Within the hydrological cycle there are several areas where water can be considered to be stored, most notably soil moisture, groundwater, snow and ice and, to a lesser extent, lakes and reservoirs. It is tempting to see stored water as static, but in reality there is considerable movement involved. The use of a storage term is explained in Figure 4.1 where it can be seen that there is an inflow, an outflow and a movement of water between the two. The inflow and outflow do not have to be equal over a time period; if not, then there has been a *change in storage* (ΔS). The critical point is that at all times there is some water stored, even if it is not the same water throughout a measurement period.

This definition of stored water is not perfect as it could include rivers as stored water in addition to groundwater, etc. The distinction is often made on the basis of flow rates (i.e. how quickly the water moves while in storage). There is no critical limit to say when a deep, slow river becomes a lake, and likewise there is no definition of how slow the flow has to be before becoming stored water. It relies on an intuitive judgement that slow flow rates constitute stored water.

The importance of stored water is highlighted by the fact that it is by far the largest amount of fresh

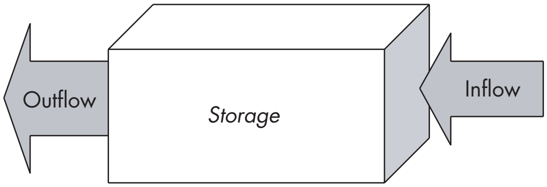

Figure 4.1 Illustration of the storage term used in the water balance equation.

water in or around planet earth (see Table 1.2, p. 6). The majority of this is either in snow and ice (particularly the polar ice caps) or groundwater. For many parts of the world groundwater is a major source of drinking water, so knowledge of amounts and replenishment rates is important for water resource management. By definition, stored water is slow moving so it is particularly prone to contamination by pollutants. The three 'Ds' of water pollution control (dilution, dispersion and degradation; see Chapter 7) all occur at slow rates in stored water, making pollution management a particular problem. When this is combined with the use of these waters for potable supply, an understanding of the hydrological processes occurring in stored water is very important.

In this chapter two major stores of water are described: water beneath the earth's surface in the unsaturated and saturated zones; and snow and ice.

WATER BENEATH THE EARTH'S SURFACE

One way of considering water beneath the earth's surface is to divide it between the saturated and unsaturated zones (see Figure 4.2). Water in the saturated zone is referred to as *groundwater* and occurs beneath a **water table**. This is also referred to as water in the **phreatic zone**.

Water in the unsaturated zone is referred to as *soil water* and occurs above a water table. This is sometimes referred to as water in the **vadose zone**.

In reality all water beneath the surface is groundwater, but it is convenient to distinguish between the saturated and unsaturated zones and maintain the terminology used by other hydrologists. As shown in Figure 4.2, there is movement

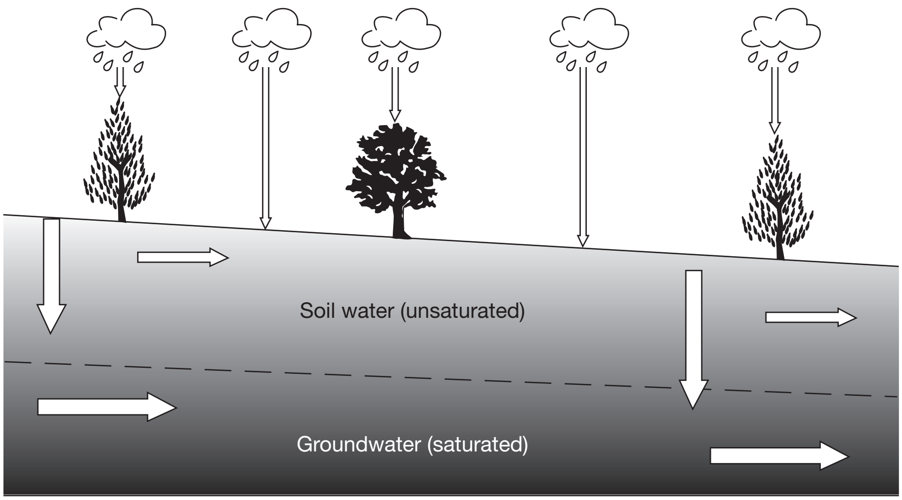

Figure 4.2 Water stored beneath the earth's surface. Rainfall infiltrates through the unsaturated zone towards the saturated zone. The broken line represents the water table, although, as the diagram indicates, this is actually a gradual transition from unsaturated to fully saturated.

of water through both vertical infiltration and horizontal flow (in reality this is a combined vector effect). It is important to realise that this occurs in both the unsaturated and saturated zones, although at a slower rate in the unsaturated.

Water in the unsaturated zone

The majority of water in the unsaturated zone is held in soil. Soil is essentially a continuum of solid particles (minerals, organic matter), water and air. Consideration of water in the soil starts with the control over how much water enters a soil during a certain time interval: the **infiltration rate**. The rate at which water enters a soil is dependent on the current water content of the soil and the ability of a soil to transmit the water.

Soil water content

Soil water content is normally expressed as a **volumetric soil moisture content** or soil moisture fraction and given the Greek symbol theta (Θ) – equation 4.1.

$$\theta = \frac{V_w}{V_t} \tag{4.1}$$

where V_w is the volume of water in a soil sample and V_t is the total volume of soil sample.

This is normally kept as unitary percentage (i.e. 1 = 100 per cent). As Θ is a volume divided by a volume it has no units, although if is sometimes denoted as m^3/m^3. Volumetric water content of a soil is effectively a depth ratio that is easily related to other equivalent depths such as rainfall and evaporation (when expressed in mm depth).

Soil water content may also be described by **gravimetric soil moisture content** (G). Gravimetric soil moisture content is the ratio of the weight of water in a soil to the overall weight of the soil. Gravimetric and volumetric soil water content can be related to each other by the soil bulk density: the density of soil in situ (equation 4.2).

$$G = \frac{M_w - M_d}{M_d} = \frac{\theta}{\rho_b \rho_w} \tag{4.2}$$

where G is the gravimetric water content (g/g); ρ_b is the bulk density of soil (g/cm^3; see equation 4.3); and ρ_w is the density of water (g/cm^3). As the density of water is close to 1 g/cm^3 it can be ignored.

Soil bulk density (ρ_b) is the ratio of the mass of dry soil to the total volume of the soil (equation 4.3).

$$\rho_b = \frac{M_d}{V_t} \tag{4.3}$$

As described above, the density of water is very close to 1 g/cm^3 (but temperature dependent; see Figure 1.3), therefore the weight of water is often assumed to be the same as the volume of water. The same cannot be said for soil: the bulk density depends on the mineralogy and packing of particles so that the volume does not equal the weight. Soil bulk density gives an indication of soil compaction with a cultivated topsoil having a value of around 1 g/cm^3 and a compacted subsoil being as high as 1.6 g/cm^3 (McLaren and Cameron, 1996). It is important to note that because of this, gravimetric soil moisture content is not the same as volumetric soil moisture content, and care must be taken in distinguishing between them as they are not interchangeable terms.

A third way of expressing soil water content is as a percentage of saturated. **Saturated water content** is the maximum amount of water that the soil can hold. Soil water content as percentage of saturated is a useful method of telling how wet the soil actually is.

Porosity (ϵ; cm^3/cm^3) is another important soil water property. It is the fraction of pore space in the total volume of soil (equation 4.4).

$$\varepsilon = \frac{V_p}{V_t} = 1 - \frac{\rho_b}{\rho_p} \tag{4.4}$$

where V_p is the volume of pores (cm^3); and ρ_p is the density of soil particles (g/cm^3).

In theory, water can fill all of the pores in a soil; therefore porosity is the maximum potential volumetric water content. In practice the volumetric soil moisture seldom reaches the porosity value and if it does gravity acts on the water to force drainage through the profile that quickly drops moisture levels back below porosity. **Field capacity** is the stable point of saturation after rapid drainage.

Other terms used in the description of soil moisture content are **soil moisture deficit** and **wilting point**. *Soil moisture deficit* is the amount of water required (in mm depth) to fill the soil up to field capacity. This is an important hydrological parameter as it is often assumed that all rainfall infiltrates into a soil until the moisture content reaches field capacity. The soil moisture deficit gives an indication of how much rain is required before saturation, and therefore when overland flow may occur (see Chapter 5). *Wilting point* is a term derived from agriculture and refers to the soil water content when plants start to die back (wilt). This is significant for hydrology as beyond this point the plants will no longer transpire.

Ability to transmit water

The ability of a soil to transmit water is dependent on the pore sizes within it and most importantly on the connections between pores. Pores can be classified according to size or function (McLaren and Cameron, 1996). Macropores are defined as pores greater than 30 μm (microns) in diameter but can also be defined by their drainage characteristic (the amount of pressure required to remove water from the pore). A well-structured soil consists of stable aggregates with a wide range of pore sizes within and between the aggregates. In this case macropores may make up at least 10 per cent of this soil volume. This structure provides numerous interconnected pathways for the flow of water with a wide range of velocities. In less well-structured soils, biological activity (e.g. roots and worms) can produce macropores that provide flowpaths for water that are largely separated from the main soil matrix (Clothier *et al.*, 1998). These are essentially two different types of macropores: those large pores within the soil matrix; and those that are essentially separated from the matrix.

The measure of a soil's ability to transmit water is **hydraulic conductivity**. When the soil is wet, water flows through the soil at a rate controlled by the saturated hydraulic conductivity (K_{sat}). The rate of flow can be described by Darcy's law, which will be explained to a fuller extent later in this chapter in the section on groundwater. When the soil is unsaturated it transmits water at a much slower rate, which is described by the Richards approximation of Darcy's law. This links the flow rate to the soil water content in a logarithmic manner so that a small change in water content can lead to a rapid increase in flow rate.

Infiltration rate

The rate at which water infiltrates the soil is not constant. Generally, water initially infiltrates at a faster rate and slows down with time (see Figure 4.3). When the infiltration rate slows down to a steady level (where the curve flattens off in Figure 4.3) the **infiltration capacity** has been reached. This is the rate of infiltration when the soil is fully saturated. The terminology of infiltration capacity is misleading as it suggests a capacity value rather than a rate. In fact infiltration capacity is the infiltration rate when the water is filled to capacity with water.

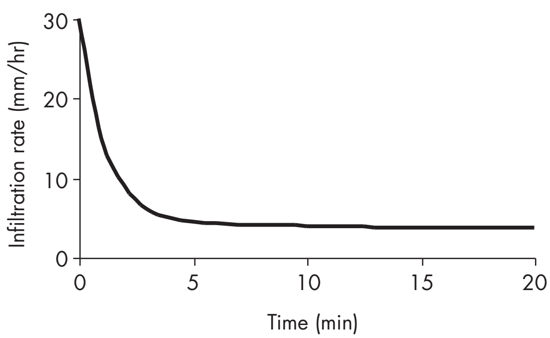

Figure 4.3 Typical infiltration curve.

Infiltration capacity is sometimes referred to as the saturated hydraulic conductivity. This is not absolutely true as the measurement is dependent on the amount of water that may be ponded on the surface creating a high hydraulic head. Saturated hydraulic conductivity should be independent of this ponded head of water. There are conditions when infiltration capacity equals saturated hydraulic conductivity, but this is not always the case.

The curve shown in Figure 4.3 is sometimes called the Philip curve, after Philip (1957) who built upon the pioneering work of Horton (1933) and provided sound theory for the infiltration of water.

The main force driving infiltration is gravity, but it may not be the only force. When soil is very dry it exerts a pulling force (soil suction; see p. 6) that will suck the infiltrating water towards the drier area. With both of these forces the infiltrating water moves down through the soil profile in a wetting front. The wetting front is three-dimensional, as the water moves outwards as well as vertically down.

The shape of the curve in Figure 4.3 is related to the speed at which the wetting front is moving. It slows down the further it gets away from the surface as it takes longer for the water at the surface to feed the front (and as the front increases in size).

Capillary forces

It is obvious to any observer that there must be forces acting against the gravitational force driving infiltration. If there were not counteractive forces all water would drain straight through to the water table leaving no water in the unsaturated zone. The counteractions are referred to as **capillary forces**. Capillary forces are actually a combination of two effects: surface tension and adsorption.

The surface tension of water is caused by the molecules in liquid water having a stronger attraction to one another than to water molecules in the air (vapour). This is due to the hydrogen bonding of water molecules described in Chapter 1. Surface tension prevents the free drainage of water from small pores within a soil by creating a force to keep the water molecules together rather than allowing them to be drawn apart. Equally, surface tension creates a force to counter the removal of water by evaporation from within the soil.

Adsorption is the force exerted through an electrostatic attraction between the faces of soil particles and water molecules. Essentially, through adsorption the water is able to stick to the surface of soil particles and not be drained away through gravity. In Chapter 1 it was pointed out that the dipolar nature of water molecules leads to hydrogen bonding (and hence surface tension). Equally the dipolar shape of water molecules lends itself to adsorptive forces.

Soil suction

Combined together, adsorption and surface tension make up the capillary force. The strength of that force is referred to as the **soil suction** or **soil moisture tension**. This reflects the concept that the capillary forces are sucking to hold onto the water and the water is under tension to keep in place. The strength of the soil suction is dependent on the amount of water present and the pore size distribution within the soil. Because of this relationship it is possible to find out pore size distribution characteristics of a soil by looking at how the soil moisture content changes at a given soil suction. This derives a *suction–moisture* or **soil moisture characteristic curve**.

A **suction–moisture curve** (see Figure 4.4) may be derived for a soil sample using a pressure plate apparatus. This uses a pressure chamber to increase the air pressure surrounding a soil sample and force water out of pores and through a ceramic plate at its base. When no more water can be forced out then it is assumed that the capillary forces (i.e. the soil suction) equal the air pressure and the sample can be weighed to measure the moisture content. By steadily increasing the air pressure between soil moisture measurements a suction–moisture curve can be derived. This can be interpreted to give important information on soil pore sizes and is also important for deriving an unsaturated hydraulic

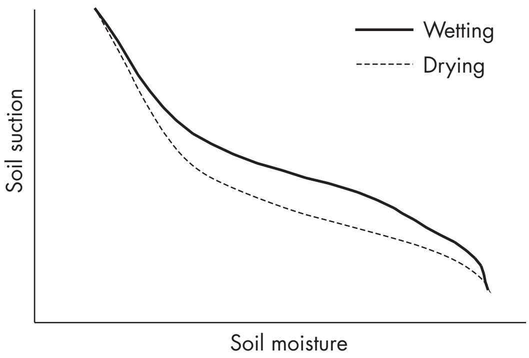

Figure 4.4 A generalised suction–moisture (or soil characteristic) curve for a soil. The two lines show the difference in measurements obtained through a wetting or drying measurement route (hysteresis).

conductivity value for a given soil moisture (Klute, 1986).

There is a major problem in interpreting suction–moisture curves, namely **hysteresis**. In short, the water content at a given soil suction depends on whether the soil is being wetted or dried. There is a different shaped curve for wetting soils than for drying ones, a fact that can be related to the way that water enters and leaves pores. It takes a larger force for air to exit a narrow pore neck (e.g. when it is drying out) than for water to enter (wetting). Care must be taken in interpreting a suction–moisture curve, as the method of measurement may have a large influence on the overall shape.

Water in the saturated zone

Once water has infiltrated through the unsaturated zone it reaches the water table and becomes groundwater. This water moves slowly and is not available for evaporation (except through transpiration in deep-rooted plants), consequently it has a long residence time. This may be so long as to provide groundwater reserves available from more pluvial (i.e. greater precipitation) times. This can be seen in the Middle East where Saudi Arabia is able to draw on extensive 'fossil water' reserves. However, it would be wrong to think that all groundwater moves slowly; it is common to see substantial movement of the water and regular replenishment during wetter months. In limestone areas the groundwater can move as underground rivers, although it may take a long time for the water to reach these conduits. In terms of surface hydrology groundwater plays an important part in sustaining streamflows during summer months.

The terminology surrounding groundwater is considerable and will not be covered in any great depth here. The emphasis is on explaining the major areas of groundwater hydrology without great detail. There are numerous texts dealing with groundwater hydrology as a separate subject, e.g. Freeze and Cherry (1979) and Price (1996).

Aquifers and aquitards

An **aquifer** is a layer of unconsolidated or consolidated rock that is able to transmit and store enough water for extraction. Aquifers range in geology from unconsolidated gravels such as the Ogallala aquifer in the USA (see Chapter 8) to distinct geological formations (e.g. chalk underlying London and much of south-east England). An **aquitard** is a geological formation that transmits water at a much slower rate than the aquifer. This is an oddly loose definition, but reflects the fact that an aquitard only becomes so relative to an aquifer. To borrow from a popular aphorism, 'one man's aquifer is another man's aquitard'. The aquitard becomes so because it is confining the flow over an aquifer. In another place the same geological formation may be considered an aquifer. The term **aquifuge** is sometimes used to refer to a totally impermeable rock formation (i.e. it could never be considered an aquifer).

There are two forms of aquifer that can be seen: confined and unconfined. A *confined aquifer* has a flow boundary (aquitard) above and below it that constricts the flow of water into a confined area (see Figure 4.5). Geological formations are the most common form of confined aquifers, and as they often occur as layers the flow of water is restricted in the vertical dimension but not in the horizontal. Water

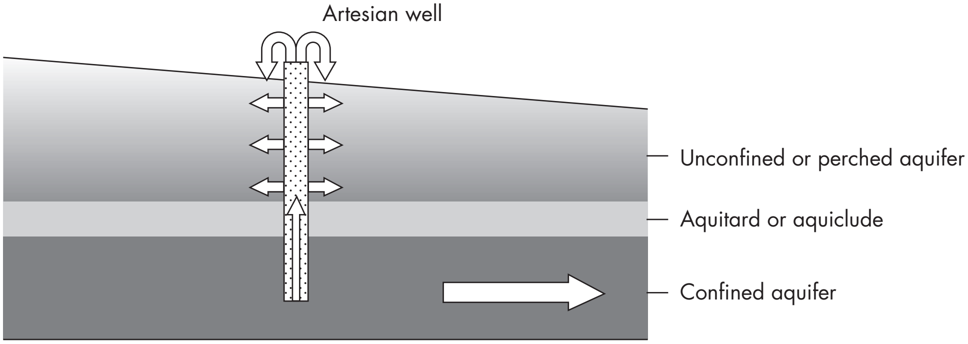

Figure 4.5 A confined aquifer. The height of water in the well will depend on the amount of pressure within the confined aquifer.

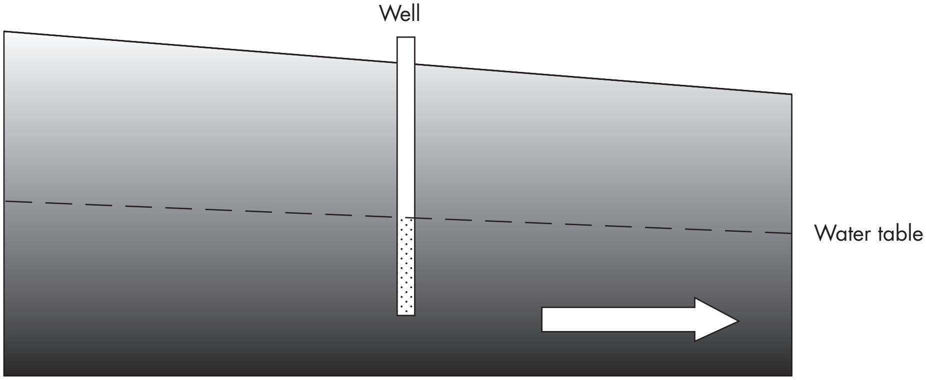

Figure 4.6 An unconfined aquifer. The water level in the well is at the water table.

within a confined aquifer is normally under pressure and if intersected by a borehole will rise up higher than the constricted boundary. If the water reaches the earth's surface it is referred to as an **artesian well**. The level that water rises up to from a confined aquifer is dependent on the amount of fall (or hydraulic head) occurring within the aquifer. This is analogous to a hose pipe acting as a syphon. If the syphon has a long vertical fall between entry of the water and exit then water will exit the hose pipe at a high velocity (i.e. under great pressure). If there is only a short vertical fall there is far less hydraulic head and the water exits at a much slower velocity. To continue the analogy further: if you could imagine that the end of the hose pipe was blocked off but that you punctured the hose, then you would expect a jet of water to shoot upwards. This jet is analogous to an artesian well.

An *unconfined aquifer* has no boundary above it and therefore the water table is free to rise and fall dependent on the amount of water contained in the aquifer (see Figure 4.6). The lower boundary of the aquifer may be impervious but it is the upper boundary, or water table, that is unconfined and may intersect the surface. It is possible to have a **perched water table** or *perched aquifer* (see Figure 4.5) where an impermeable layer prevents the infiltration of water down to the regional water table. Perched water tables may be temporary features reflecting variable hydraulic conductivities within the soil and rock, or they can be permanent features reflecting the overall geology.

Groundwater flow

The movement of water within the saturated zone is described by Darcy's law (equation 4.5). Henri Darcy was a nineteenth-century French engineer concerned with the water supply for Dijon in France. The majority of water for Dijon is aquifer fed and Darcy began a series of observations on the characteristics of flow through sand. He observed that the 'rate of flow of water through a porous medium was proportional to the hydraulic gradient' (Darcy, 1856). There are many different ways of formulating Darcy's law, but the most common and easily understood is shown in equation 4.5.

$$Q = -K_{sat} \cdot A \cdot \frac{dh}{dx} \tag{4.5}$$

The **discharge** (Q) from an aquifer equals the saturated hydraulic conductivity (K_{sat}) multiplied by the cross-sectional area (A) multiplied by the hydraulic gradient (dh/dx). The negative sign is convention based on where you measure the hydraulic gradient from (i.e. a fall in gradient is negative).

The h term in the hydraulic gradient includes both the elevation and pressure head. In an unconfined aquifer it can be assumed that the hydraulic gradient is equal to the drop in height of water table over a horizontal distance (i.e. the elevation head). In a confined aquifer it is the drop in phreatic surface (i.e the level that water in boreholes reaches given the pressure the water is under) over a horizontal distance. The h term then includes a pressure head.

Darcy's law is an empirical law (i.e. based on experimental observation) that appears to hold under many different situations and spatial scales. It underlies most of groundwater hydrology and is very important for the management of groundwater resources. It is the term 'hydraulic conductivity' (K_{sat}) that is so important. This is the ability of a porous medium to transmit water. This can be related to the size of pores within the soil or rock and the interconnectivity between these pores. Table 4.1 shows some common values of hydraulic conductivities for soils, in addition to their porosity values.

Table 4.1 Soil hydrological properties for selected soil types

Soil type	*Saturated hydraulic conductivity (cm/hr)*	*Porosity*
Sandy loam	2.59	0.45
Silt loam	0.68	0.50
Clay loam	0.23	0.46
Clay	0.06	0.475

Source: Derived from measurements of different soil types in the USA. From Rawls *et al.* (1982)

One of the major difficulties in applying Darcy's law is that hydraulic conductivities vary spatially at both micro and macro scales. Although K_{sat} can be measured from a small sample in the laboratory (Klute and Dirksen, 1986), in the management of water resources it is more common from larger-scale well-pumping tests (see Freeze and Cherry (1979) for more details). The well-pumping test gives a spatially averaged K_{sat} value at the scale of interest to those concerned with water resources.

The relationship between groundwater and surface water

It is traditional to think of groundwater sustaining streamflows during the summer months, which indeed it often does. However, the interaction between groundwater and streamflow is complex and depends very much on local circumstances. Water naturally moves towards areas where faster flow is available and consequently can be drawn upwards towards a stream. This is the case in dry environments but is dependent on there being an

Case study

METHOD – AGEING GROUNDWATER

An important piece of information for somebody managing groundwater resources is the age of the water contained in the aquifer. This information will give an idea of how quickly any contaminated water may move towards an extraction zone, or how long ago the contamination occurred. Darcy's Law gives an indication of the possible flow rates within an aquifer but the measurement is at too small a scale to be scaled up to estimate how long it has taken for water to reach a certain position. Frequently there is little idea of where the water has actually come from, so even if you could estimate water velocities you don't know the distance travelled and therefore can't estimate age. In order to overcome this problem groundwater scientists use the chemistry of different substances dissolved within the water to estimate its age.

Carbon dating is a common technique for the dating of terrestrial deposits but is problematic for young groundwater, since it is only accurate when the sample is more than 1,000 years old. Groundwater is frequently more than 1,000 years old so it is possible to use the technique of carbon dating, looking at the rate of decay of ^{14}C in dissolved organic carbon. Another form of carbon dating looks for ^{14}C resulting from the testing of thermo-nuclear weaponry.

Another dating method, particularly for younger groundwater, is to look for concentrations of materials that we know have been added to the atmosphere by humans. Fortunately for groundwater dating, humans have been very good at polluting the atmosphere with substances that are then dissolved in precipitation. Figure 4.7 shows the concentration of four gases that have been added to the atmosphere in differing amounts. The relative concentrations of these gases in water samples give an estimate for the average age of the groundwater. Tritium, a radioactive isotope of hydrogen with 3 neutrons, was added to the atmosphere in large quantities through explosion of hydrogen bombs, particularly in the 1960s and 1970s. Tritium concentrations reached a peak in 1963 and have since declined to almost background concentrations. Tritium has a radioactive half-life of 12.3 years. Chlorofluorocarbon (CFC) compounds were used in aerosols and refrigeration from the 1940s until their banning in the 1990s. CFC-11 concentration has slowly declined since about 1993, while CFC-12 concentration is still increasing, but at a much slower rate than before 1990. Sulphur hexafluoride (SF_6) is used for cooling and insulation, particularly in electronics.

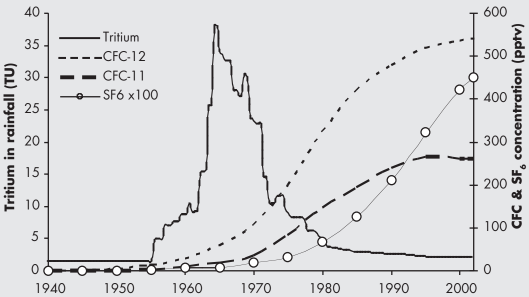

Figure 4.7 Tritium concentrations in rainfall, CFC and SF_6 concentrations in the atmosphere 1940–2002. Tritium units (TU) are 1 tritium atom in 1,018 hydrogen atoms. CFC and SF_6 are in parts per trillion by volume (pptv).

Source: Figure redrawn from Stewart *et al*. (2007)

Another dating method is to look at the ratio of two isotopes of oxygen and/or two isotopes of hydrogen found in water molecules. When water in the atmosphere condenses to form rain there is a preferential concentration of heavy isotopes of hydrogen and oxygen in the water molecules. The heavy isotope of hydrogen is deuterium (1 neutron) and the heavy isotope of oxygen is ^{18}O.

The colder the temperature is at time of condensation the more enriched with deuterium and ^{18}O the water sample will be. In climates with distinct seasons the amount of deuterium and ^{18}O in rainfall samples will vary according to seasons. By taking a series of water samples from rainfall and groundwater a comparison can be made between the time series. If the groundwater shows considerable variation in deuterium and/or ^{18}O concentrations then it is relatively recent rainfall. If there is very little variation then it is assumed that the groundwater is a mixture of rainfall from both summers and winters in the past and is therefore older. This is demonstrated in Figure 4.8. The technique of looking at oxygen and hydrogen isotopes is particularly common as a way of determining whether water in a stream is new (i.e. recently derived from rainfall) or old (has been resident in groundwater for sometime).

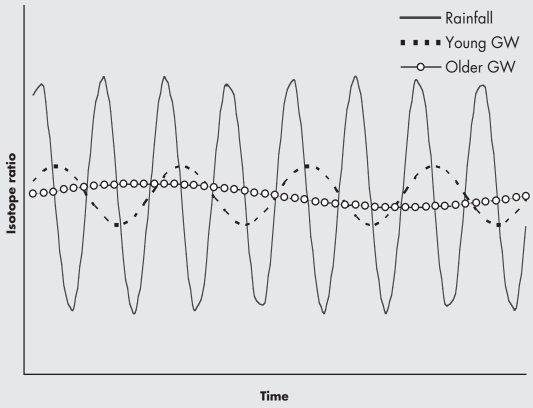

Figure 4.8 Changing ratios of isotopes of oxygen and hydrogen with time in a seasonal climate. Rainfall is heavily influenced by temperature and shows large variation between seasons. The older the groundwater the more dampened down the time series.

By measuring the concentrations of contaminants like tritium or the ratio of isotopes of oxygen an estimate of the age of groundwater can be made. The different way that water moves through the unsaturated and saturated zones of catchments means that groundwater and streamflows contain water with different residence times. The water in a sample does not have a discrete age, but has a distribution of ages. This distribution is described by a conceptual flow or mixing model, which reflects the average conditions in the catchment (Stewart *et al.*, 2007). Maloszewski and Zuber (1982) provide an extensive review of these mixing models; in short they account for the size of groundwater reservoir, the concentration of contaminant and compute the likely time of residence within a well mixed groundwater reservoir.

Using the techniques outlined here, the average age of groundwater or streamwater can be derived. Studies of groundwater age frequently use a combination of the different techniques to derive an average residence time of water in a catchment, or the groundwater age. It is important to realise that it is an average residence time, not absolute. The water contained in the groundwater reservoir will be a mixture of water that has infiltrated rapidly and some that moved very slowly through the unsaturated zone.

unconfined aquifer near to the surface. If this is not the case then the stream may be contributing water to the ground through infiltration. Figure 4.9 shows two different circumstances of interaction between the groundwater and stream. In Figure 4.9(a) the groundwater is contributing water to the streamflow as the water table is high. In Figure 4.9(b) the water table is low and the stream is contributing water to the groundwater. This is commonly the case where the main river source may be mountains a considerable distance away and the river flows over an alluvial plain with the regional groundwater table considerably deeper than stream level. The interaction between groundwater and streamflow is discussed further in Chapter 5, especially with respect to stormflows.

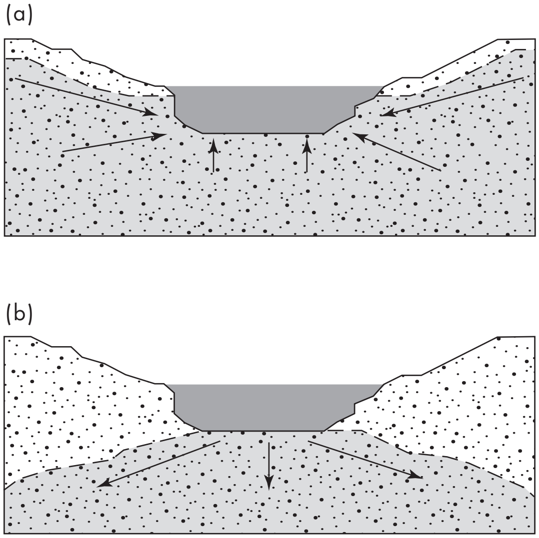

Figure 4.9 The interactions between a river and the groundwater. In (a) the groundwater is contributing to the stream, while in (b) the opposite is occurring.

Measuring water beneath the surface

Measurement of soil water

Gravimetric method: The simplest and most accurate means for the measurement of soil water is using the gravimetric method. This involves taking a soil sample, weighing it wet, drying in an oven and then weighing it dry. Standard practice for the drying of soils is 24 hours at 105°C (Gardner, 1986). The difference between the wet and dry weights tells you how wet the soil was. If it is volumetric soil moisture content that is required then you must take a sample of known volume. This is commonly done using an undisturbed soil sampler and utilising equation 4.6.

$$\theta = \frac{V_w}{V_t} \approx \frac{M_w - M_d}{V_t} \tag{4.6}$$

Where Θ is the volumetric water content (cm^3/cm^3); V_w is the volume of water (cm^3); V_t is the total volume of soil (cm^3); M_w is the mass of water (g); M_d is the mass of dry soil (g). The density of water is close to 1 g/cm^3 so the weight of water can be assumed to equal the volume of water; hence the ≈ symbol in the equation above.

Gravimetric analysis is simple and accurate but does have several drawbacks. Most notable of these is that it is a destructive sampling method and therefore it cannot be repeated on the same soil sample. This may be a problem where there is a requirement for long-term monitoring of soil moisture. In this case a non-destructive moisture-sampling method is required. There are three methods that fit this bill, but they are indirect estimates of soil moisture rather than direct measurements as they rely on measuring other properties of soil in water. The three methods are: **neutron probes**, *electrical resistance blocks* and **time domain reflectometry**. All of these can give good results for monitoring soil moisture content, but are indirect and require calibration against the gravimetric technique.

Neutron probe: A neutron probe has a radioactive source that is lowered into an augured hole; normally the hole is kept in place as a permanent access tube using aluminium tubing. The radioactive source emits fast (or high energy) neutrons that collide with soil and water particles. The fast neutrons are very similar in size to a hydrogen ion (H^+ formed in the disassociation of the water molecule) so that when they collide the fast neutron slows down and the hydrogen ion speeds up. In contrast, when a fast neutron collides with a much larger soil particle it bounces off with very little loss of momentum. The analogy can be drawn to a pool table. When the cue ball (i.e. a fast neutron) collides with a coloured pool ball (i.e. a water particle) they both move off at similar speeds, the cue ball has slowed down and the coloured ball speeded up. In contrast, if the cue ball hits the cushion on the edge of a pool table (i.e. a soil particle) it bounces off with very little loss of speed. Consequently the more water there is in a soil the more fast neutrons would slow down to become 'slow neutrons'. A neutron probe counts the number of slow neutrons returning towards the radioactive tip, and this can be related to the soil moisture content. The neutron probe readings need to be calibrated against samples of soil with known moisture contents. This is often done by using gravimetric analysis on the samples collected while the access tubes are being put in place. It is also possible to calibrate the probe using reconstituted soil in a drum or similar vessel. It is important that the calibration occurs on the soil actually being measured, since the fast:slow neutron ratio will vary according to mineralogy of the soil.

Although the neutron probe is essentially non-destructive in its measurement of soil water content, it is not continuous. There is a requirement for an operator to spend time in the field taking measurements at set intervals. This may present difficulties in the long-term monitoring of soil moisture. Another difficulty with a neutron probe is that the neutrons emitted from the radioactive tip move outwards in a spherical shape. When the probe tip is near the surface some of the sphere of neutrons will leave the soil and enter the atmosphere, distorting the reading of returning slow neutrons. A very careful calibration has to take place for near-surface readings and caution must be exercised interpreting these results. This is unfortunate as it is often the near-surface soil moisture content that is of greatest importance. Although neutron probes are reliable instruments for the monitoring of soil moisture the cost of the instruments, difficulties over installing access tubes (Figure 4.10), calibration problems and the near-surface problem have meant that they have seldom been used outside a research environment.

Figure 4.10 A neutron probe sitting on an access tube. The black cable extends down into the tube with the source of fast neutrons (and counter) at the tip.

Electrical resistance blocks: Electrical resistance blocks use a measurement of electrical resistance to infer the water content of a soil. As water is a conductor of electricity it is reasonable to assume that the more water there is in a soil the lower the electrical resistance, or conversely, the higher the electrical conductivity. For this instrumentation two small blocks of gypsum are inserted into the soil and a continuous measurement of electrical resistance between the blocks recorded. The measure of electrical resistance can be calibrated against gravimetric analysis of soil moisture. The continuity of measurement in electrical resistance blocks is a great advantage of the method, but there are several problems in interpreting the data. The main difficulty is that the conductivity of the water is dependent on the amount of dissolved ions contained within it. If this varies, say through the application of fertiliser, then the electrical resistance will decrease in a manner unrelated to the amount of water present. The second major difficulty is that the gypsum blocks deteriorate with time so that their electrical conductivity alters. This makes for a gradually changing signal, requiring constant recalibration. The ideal situation for the use of electrical resistance blocks is where they do not sit in wet soil for long periods and the water moving through the soil is of relatively constant dissolved solids load. An example of this type of situation is in sand dunes, but these are not particularly representative of general land use.

Time domain reflectometry: Time domain reflectometry (TDR) is a relatively new soil moisture measurement technique. The principle of measurement is that as a wave of electromagnetic energy is passed through a soil the wave properties will alter. The

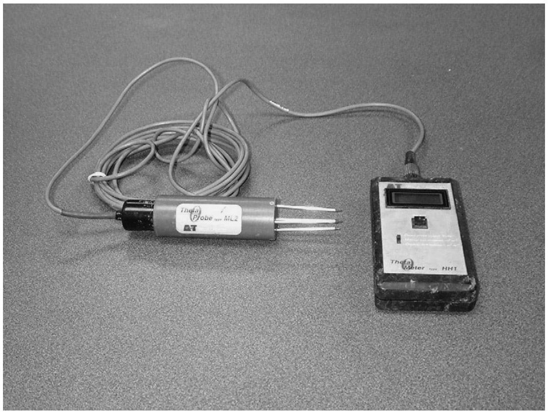

Figure 4.11 The Theta probe (manufactured by Delta-T devices). An example of a small, time domain reflectometry instrument used to measure soil moisture content in the field. The metal spikes are pushed into the soil and the moisture level surrounding them is measured.

way that these wave properties alter will vary, dependent on the water content of the soil. TDR measures the properties of microwaves as they are passed through a soil and relates this to the soil moisture content. Although this sounds relatively simple it is a complicated technique that requires detailed electronic technology. Up until the late 1990s this had restricted the usage of TDR to laboratory experiments but there are now soil moisture probes available that are small, robust and reliable in a field situation. An example of this is the Theta probe shown in Figure 4.11.

Tensiometers

A **tensiometer** is used to measure the soil suction pressure or soil moisture tension. This is the force exerted by capillary forces and it increases as the soil dries out. A tensiometer is a small ceramic cup on the end of a sealed tube of water. The dry soil attempts to suck the water from the water-filled tube through the ceramic cup. At the top of the tube a diaphragm measures the pressure exerted by this suction. Tensiometers also have the ability to measure a positive pressure when the soil water is held under pressure (e.g. during a rising water table). The units of soil suction are Pascals, the SI units for pressure.

Piezometers and wells

Most of the techniques described in this section have been for the measurement of water in the unsaturated zone. The main measurement techniques for water in the saturated zone are through **piezometers** and **wells**. Both of these measure the height of the water table but in slightly different ways. A *piezometer* is a tube with holes at the base that is placed at depth within a soil or rock mantle. The height of water recorded in the piezometer is thus a record of the pressure exerted by the water at the base of the tubing. A *well* has permeable sides all the way up the tubing so that water can enter or exit from anywhere up the column. Wells are commonly used for water extraction and monitoring the water table in unconfined aquifers. Piezometers are used to measure the water pressure at different depths in the aquifer.

Measurement of infiltration rate

Infiltration rate is measured by recording the rate at which water enters the soil. There are numerous methods available to do this, the simplest being a ring **infiltrometer** (see Figure 4.12). A solid ring is pushed into the ground and a pond of water sits on the soil (within the ring). This pond of water is kept at a steady level by a reservoir held above the ring. Recordings of the level of water in the reservoir (with time) give a record of the infiltration rate. To turn the infiltration volume into an infiltration depth the volume of water needs to be divided by the cross-sectional area of the ring.

A simple ring infiltrometer provides a measure of the ponded infiltration rate, but there are several associated problems. The first is that by using a single ring a large amount of water may escape around the sides of the ring, giving higher readings than would be obtained from a completely saturated

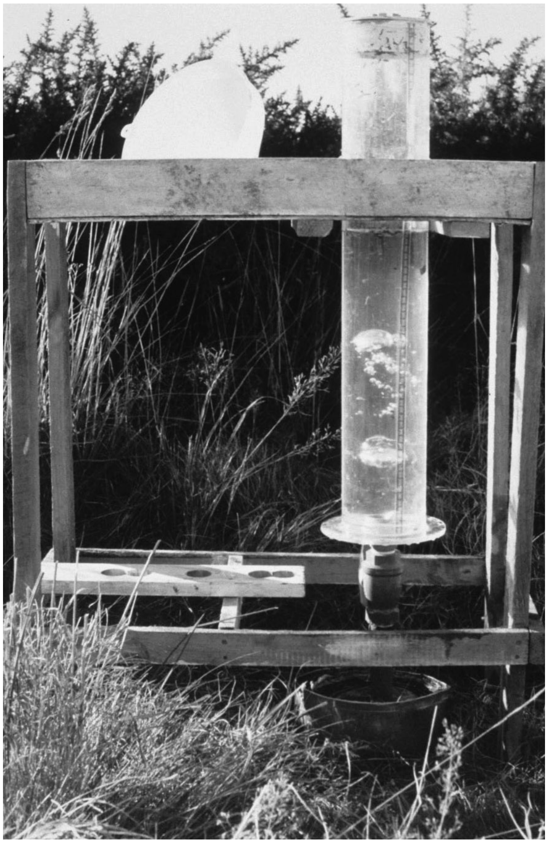

Figure 4.12 A single ring infiltrometer. The ring has been placed on the ground and a pond of water is maintained in the ring by the reservoir above. A bubble of air is moving up the reservoir as the water level in the pond has dropped below the bottom of the reservoir. A reading of water volume in the reservoir is taken and the time recorded.

surface. To overcome this a double ring infiltrometer is sometimes used. With this, a second wider ring is placed around the first and filled with water so that the area surrounding the measured ring is continually wet. The second problem is that ponded infiltration is a relatively rare event across a catchment. It is more common for rainfall to infiltrate directly without causing a pond to form on the surface. To overcome this a rainfall simulator may be used to provide the infiltrating water.

Estimating water beneath the surface

In the previous section it was stated that several of the methods listed were indirect measurement methods or estimation techniques. They certainly do not measure soil moisture content directly, but they have a good degree of accuracy and are good measures of soil moisture, albeit in a surrogate form. Estimating the amount of water beneath the surface can also be carried out using either numerical modelling or remote sensing techniques. The main groundwater modelling techniques focus on the movement of water in the subsurface zone, using different forms of Darcy's law. There are also models of soil water balance that rely on calculating inflows (infiltration from rainfall) and outflows (seepage and evaporation) to derive a soil water storage value for a given time and space.

A less reliable technique for the estimation of soil moisture is through remote sensing. There are three satellite remote sensing techniques of relevance to soil moisture assessment: thermal imagery, passive microwave and active microwave. All of these techniques sense the soil moisture content at the very near surface (i.e. within the top 5 cm), which is a major restriction on their application. However, this is an important area in the generation of runoff (see Chapter 5) and is still worthy of measurement.

Thermal imagery

The high heat capacity of water means that it has considerable effect on the emission of thermal infrared signals from the earth. These can be detected by satellites and an inference made about how wet the soil is. This is especially so if two images of the same scene can be compared in order to derive a relative wetness. Satellite platforms like LANDSAT and SPOT are able to use this technique at spatial resolutions of around 10–30 m. The major difficulty with this technique is that it relies on a lack of cloud cover over the site of investigation, something that is not easy to guarantee, especially in hydrologically active (i.e. wet) areas.

Passive microwave

The earth surface emits microwaves at a very low level that can be detected by satellites. These are referred to as passive microwaves, in the sense that the earth emits them regardless of whether the satellite is present or not. The amount of water present on or above the surface of the earth affects the passive microwave signal emitted (it is a lower signal the more water there is present). The SSM/I satellite platform is able to measure passive microwaves, but only at a very coarse spatial resolution ($\approx 100 \times 100$km). This is the major drawback of this technique at present.

Active microwave

Active microwaves are emitted from a satellite and the strength of returning signal measured. This is a complex radar system and has only been available on satellites since the early 1990s. The strength of microwave backscatter is primarily dependent on two factors: the soil wetness and the surface roughness. Where soil roughness is well known and there is little or no vegetation cover the radar backscatter has been well correlated with surface moisture (Griffiths and Wooding, 1996; Kelly *et al*., 2003). This technique offers hope for future estimation of soil moisture from satellites, but there is still much research to be done in understanding the role vegetation plays in affecting radar backscatter.

The great advantage of any remote sensing technique is that it automatically samples over a wide spatial area. The satellite measures the electromagnetic radiation within each pixel; this is an average value for the whole area, rather than a point measurement that might be expected from normal soil moisture measurements. The question that needs to be answered before satellite remote sensing is widely accepted in hydrology is whether the enhanced spatial distribution of measurement is sufficient to overcome the undoubted limitations of direct soil moisture measurement.

Case study

REMOTE SENSING OF SOIL MOISTURE AS A REPLACEMENT FOR FIELDWORK

Remote sensing of soil moisture may offer a way of deriving important hydrological information without intensive, and costly, fieldwork programmes. Grayson *et al*. (1992) suggest that this could be used to set the initial conditions for hydrological modelling, normally a huge logistical task. One of the major difficulties in this is that the accuracy of information derived from satellite remote sensing is not high enough for use in hydrological modelling. As a counter to this it can be argued that the spatial discretisation offered by remote sensing measurements is far better than that available through traditional field measurement techniques.

In an attempt to reconcile these differences Davie *et al*. (2001) intensively monitored a 15-hectare field in eastern England and then analysed the satellite-derived, active microwave backscatter for the same period. The field programme consisted of measuring surface soil moisture (gravimetric method) at three different scales in an attempt to spatially characterise the soil moisture. The three scales consisted of: (a) thirteen samples taken 1 m apart; (b) lines of samples 30 m apart; and (c) the lines were approximately 100 m apart. The satellite data was from the European Remote Sensing Satellite (ERS) using **Synthetic Aperture Radar** (SAR).

Analysis of the field data showed great variation in the surface soil moisture and a difference in measurement depending on the scale of measurement (see Figure 4.13). It is evident from Figure 4.13 that the point measurements of soil moisture are highly variable and that many measurements of soil moisture are required to try and characterise the overall field surface soil moisture.

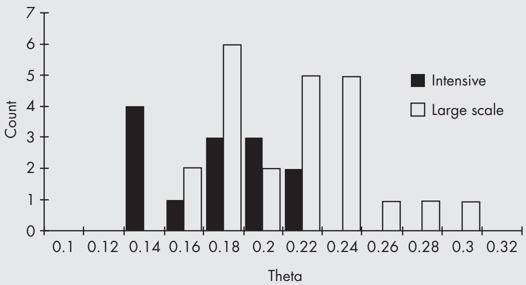

Figure 4.13 Measured surface soil moisture distributions at two different scales for a field in eastern England in October 1995.

In contrast to the directly measured soil moisture measurements, the microwave backscatter shows much less variation in response (see Figure 3, p. 330 of Davie *et al*., 2001). The most likely explanation for this is that the spatial averaging of backscatter response within a pixel (25 × 25m in this case) evens out the variations found through point measurements. This can be thought of as a positive aspect because in many cases for hydrological modelling the scale needed is larger than point measurements and the spatial integration of backscatter response may be a way of getting around highly variable point measurements. When further analysis of the backscatter response was carried out it showed that interpretation at the pixel size was meaningless and they needed to be averaged-up themselves. It was found that the smallest resolution to yield meaningful results was around 1 hectare (100 × 100 m).

As with other studies on bare fields (e.g. Griffiths and Wooding, 1996) there was a reasonably good correlation between radar backscatter and measured soil moisture. Unfortunately, this relationship is not good enough to provide more than around 70 per cent accuracy on estimations of soil moisture. In addition to this the study was carried out in conditions ideally suited for SAR interpretation (flat topography with no vegetation cover) which are far from atypical conditions encountered in hydrological investigations.

Overall it is possible to say that although the satellite remote sensing of soil moisture using SAR may offer some advantages of spatial integration of the data it is not enough to offset the inaccuracy of estimation, particularly in non-flat, vegetated catchments. There may come a time when satellite remote sensing can be used as a replacement for field measurements, but this will be in the future.

SNOW AND ICE

Snow and ice are hugely important stores of water for many countries in the world, particularly at high latitudes or where there are large mountain ranges. The gradual release of water from snow and ice, either during spring and summer or on reaching a lower elevation, makes a significant impact on the hydrology of many river systems.

In the same manner that rainfall may be intercepted by a canopy, so can snow. The difference between the two is in the mass of water held and the duration of storage (Lundberg and Halldin, 2001). The amount of intercepted snow is frequently much higher than for rainwater and it is held for much longer. This may be available for evaporation through sublimation (moving directly from a solid to a gas) or release later in snow melt. Hedstrom and Pomeroy (1998) point out that the mass of snow held by interception is controlled by the tree branching structure, leaf area and tree species. In countries such as Canada and Russia there are extensive forests in regions dominated by winter snowfall. Some studies have shown as much as 20–50 per cent of gross precipitation being intercepted and evaporated (Lundberg and Halldin, 2001). These figures indicate that a consideration of snowfall interception is critical in these regions. Lundberg and Halldin (2001) provide a recent review of measuring snow interception and modelling techniques.

The timing of snow and ice melt is critical in many river systems, but especially so in rivers that flow north towards the Arctic Circle. In this case the melting of snow and ice may occur in the headwaters of a river before it has cleared further downstream (at higher latitudes). This may lead to a build up of water behind the snow and ice further downstream – a snow and ice dam (see Figure 4.14). Beltaos (2000) estimates that the cost of damage caused by ice-jams in Canada is around $60 million dollars per annum. The Case Study on pp. 72–75 gives an example of the flow regime that may result from this type of snow melt event.

Figure 4.14 Susquehanna river ice jam and flood which destroyed the Catawissa Bridge in Pennsylvania, USA on 9 March 1904.

Source: Photo copyright of Columbia County Historical and Genealogical Society

Case study

THE MACKENZIE RIVER: DEMONSTRATING THE INFLUENCE OF SNOW AND ICE ON RIVER FLOWS

The Mackenzie river is at the end of one of the great river systems of the world. In North America the Missouri/Mississippi system drains south, the Great Lakes/Hudson system drains eastwards, and the Mackenzie river system drains northwards with its mouth in the Beaufort Sea (part of the Arctic Ocean). The Mackenzie river itself has a length of 1,800 km from its source: the Great Slave Lake. The Peace and Athabasca rivers which flow into the Great Slave Lake, and are therefore part of the total Mackenzie drainage basin, begin in the Rocky Mountains 1,000 km to the southwest. The total drainage basin is approximately 1,841,000 km^2, making it the twelfth largest in the world, and the river length is 4,250 km (see Figure 4.15).

The high latitude of the Mackenzie river makes for a large component of snow and ice melt within the annual hydrograph. This can be seen in Figure 4.16, taken at the junction of the Mackenzie and Arctic Red rivers. This gauging station is well within the Arctic circle and towards the delta of the Mackenzie river. The monthly discharge values increase dramatically from April to June when the main melt occurs and then gradually decrease to reach a minimum value during the winter months. The highest average streamflow occurs in June despite the highest precipitation occurring one

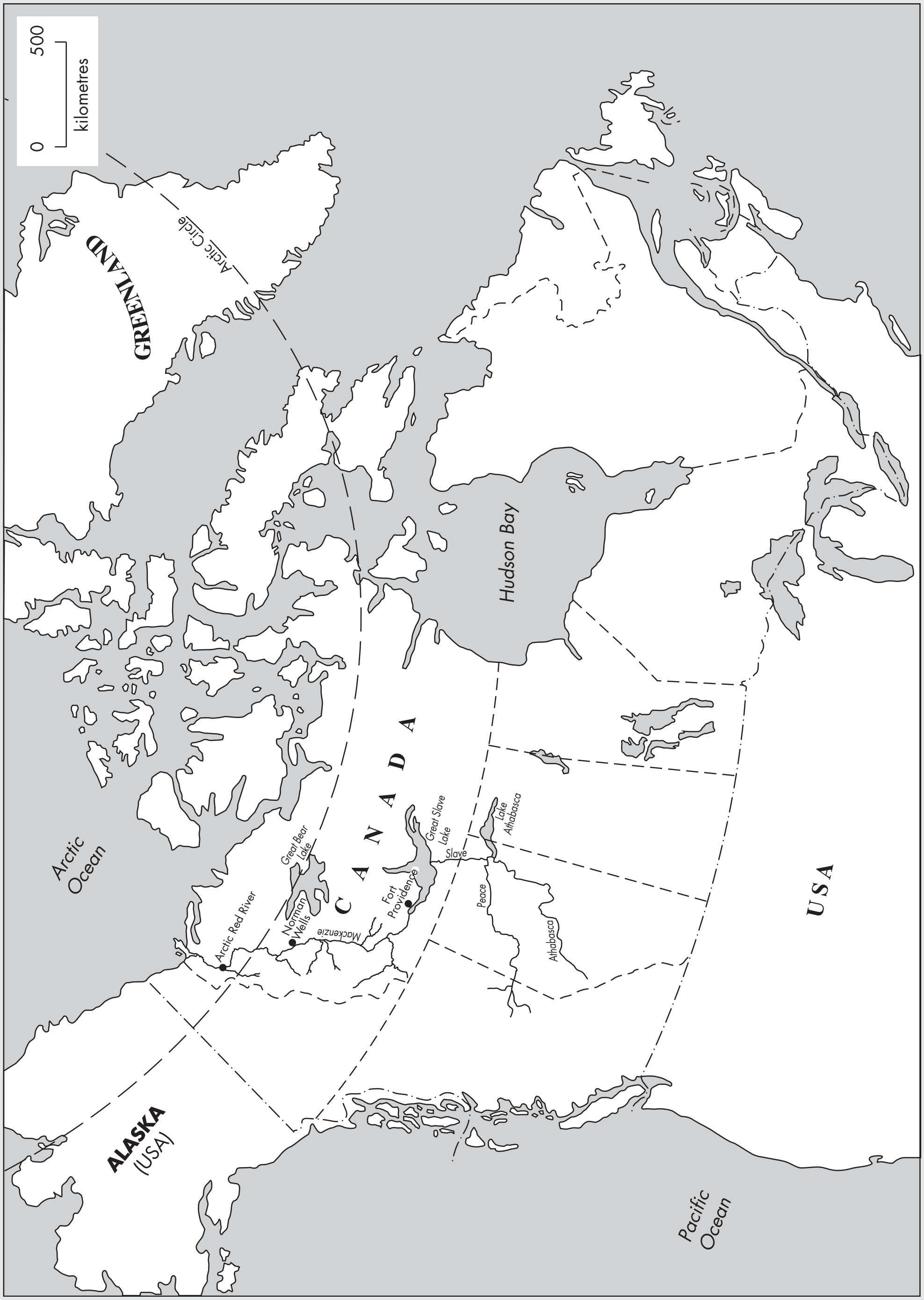

Figure 4.15 Location of the Mackenzie river in Canada.

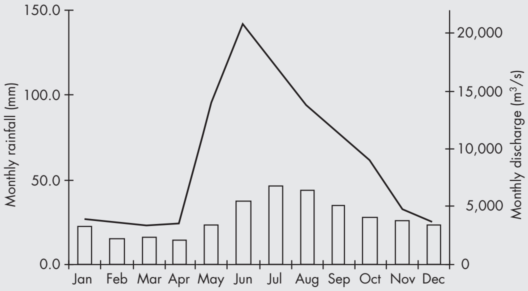

Figure 4.16 Average monthly river flow (1972–1998; line) for the Mackenzie river at the Arctic Red River gauging station (latitude 67° 27' 30'' north) and average precipitation (1950–1994) for the Mackenzie river basin (bars).

Source: Data courtesy of Environment Canada

month later in July. Overall there is very little variation in precipitation but a huge variation on riverflow. This an excellent example of the storage capability of snow and ice within a river catchment. Any water falling during the winter months is trapped in a solid form (snow or ice) and may be released only during the warmer summer months. The amount of precipitation falling during the summer months (mostly rainfall) is dwarfed by the amount of water released in the melt during May and June.

The most remarkable feature of a river system such as the Mackenzie is that the melt starts occurring in the upper reaches, sending a pulse of water down the river, before ice on the lower reaches has melted properly. This is not unique to the Mackenzie river, all the great rivers draining northwards in Europe, Asia and North America exhibit the same tendencies. If we look in detail at an individual year (Figure 4.17) you can see the difference in daily hydrographs for stations moving down the river (i.e. northwards). Table 4.2 summarises the information on latitude and flow characteristics for the Mackenzie. It is not a simple story to decipher (as is often the case in hydrology), but you can clearly see that the rise in discharge at the Arctic Red River station starts later than the Norman Wells station further

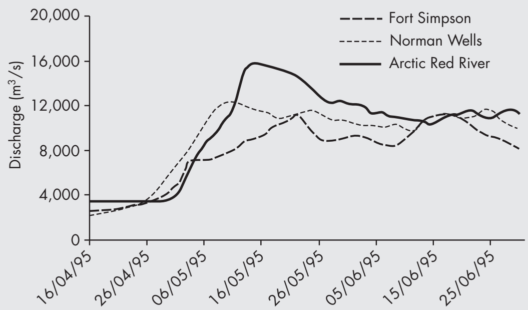

Figure 4.17 Daily river flow at three locations on the Mackenzie river from mid-April through to the end of June 1995.

Source: Data courtesy of Environment Canada

Table 4.2 Summary of latitude and hydrological characteristics for three gauging stations on the Mackenzie river

Mackenzie river gauging station	*Latitude (north)*	*Date of last ice on river (1995)*	*Date of peak discharge (1995)*
Fort Simpson	61° 52′ 7″	14 May	21 May
Norman Wells	65° 16′ 26″	18 May	10 May
Arctic Red River	67° 27′ 30″	31 May	14 May

Source: Data courtesy of Environment Canada to the south. The rise is caused by melt, but predominantly from upstream. It is also clear that for both Norman Wells and the Arctic Red River stations the highest discharge value of the year is occurring while the river is still covered in ice. This creates huge problems for the drainage of the area as the water may build up behind an ice dam. Certainly the water flowing under the ice will be moving much quicker than the ice and water mix at the surface. Plate 5 demonstrates the way that water builds up behind an ice dam, particularly where there is a constriction on either side of the river channel.

Measuring snow depth

The simplest method of measuring snow depth is the use of a core sampler. This takes a core of snow, recording its depth. The snow sample can then be melted to derive the water equivalent depth, the measurement of most importance in hydrology. The major drawbacks in using a core sampler to derive snow depth are that it is a non-continuous reading (similar to daily rainfall measurement) and the position of coring may be critical (because of snow drifting).

A second method of measuring snow depth is to use a *snow pillow*. This is a sealed plastic pillow that is normally filled with some form of anti-freeze and connected to a pressure transducer (Figure 4.18). When left out over a winter period the weight of snow on top of the pillow is recorded as an increasing pressure, which can be recalculated into a mass of snow. It is important that the snow pillow does not create an obstacle to drifting snow in its own right. To overcome this, and to lessen the impact of freezing on the pillow liquid, it is often buried under a shallow layer of soil or laid flat on the ground. When connected to a continuous logging device a snow pillow provides the best record of snow depth (and water equivalent mass).

Estimating snow cover

The main method of estimating snow cover is using satellite remote sensing. Techniques exist that give a reasonably good method of detecting the areal extent of snow cover, but it is much harder to translate this into a volume of snow (i.e. by knowing depth) or water. Optical and thermal infrared data can be used to estimate snow cover but they rely heavily on the reflective ability of snow; unfortunately, other surfaces such as clouds may also exhibit these properties (Fitzharris and McAlevey,

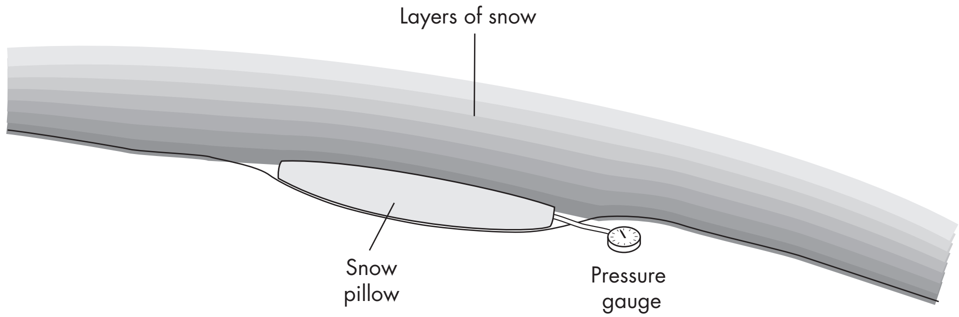

Figure 4.18 Snow pillow for measuring weight of snow above a point. The snow weight is recorded as a pressure exerted on the pillow.

1999). Microwave data offer a far better method of detecting snow cover. Passive microwaves detected by a satellite can be interpreted to give snow cover because any water (or snow) covering the surface absorbs some of the microwaves emitted by the earth surface. The greater the amount of snow the weaker is the microwave signal received by the satellite. Ranzi *et al*. (1999) have used AVHRR imagery to monitor the snowpack in an area of northern Italy and some of Switzerland and compared this to other measurement techniques. Unfortunately, the passive microwave satellite sensors currently available (e.g. AVHRR) are at an extremely coarse spatial resolution that is really only applicable at the large catchment scale. Active microwave sensing offers more hope but its usage for detecting snow cover is still being developed.

Snow melt

Of critical importance to hydrology is the timing of snow melt, as this is when the stored water is becoming available water. There are numerous models that have been developed to try and estimate the amount of snow melt that will occur. Ferguson (1999) gives a summary of recent snow melt modelling work. The models can be loosely divided into those that rely on air temperature and those that rely on the amount of radiation at a surface. The former frequently use a degree days approach, the difference between mean daily temperature and a melting threshold temperature. Although it would seem sensible to treat zero as the melting threshold temperature this is not always the case; snow will melt with the air temperature less than zero because of energy available through the soil (soil heat flux) and solar radiation. The degree day snow melt approach calibrates the amount of snow that might be expected given a certain value of degree day. Although this is useful for hydrological studies it is often difficult to calibrate the model without detailed snow melt data.

STORAGE IN THE CONTEXT OF WATER QUANTITY AND QUALITY

The storage term within the water balance equation is among the most important for consideration of water quantity and quality. Its influence on water quantity has been outlined in this chapter (e.g. snow melt contribution to flooding, groundwater contribution to streamflow – see Chapter 5). The importance of storage for water quality is mostly through the addition and removal of nutrients that occurs when water comes into contact with soil. The soil is an extremely active biological zone, whether through microbes, plants or animals. Water is an important part of the nutrient cycling that occurs within the soil zone; the presence or absence of water in a soil limits the biological productivity of a site. When water passes through soil it dissolves chemical salts (both natural and introduced to the system by humans as fertilisers). Many of these salts are used by plants and microbes as part of their respiration processes. Those that are not used may stay in the water as it moves into groundwater and possibly back into river systems. This is referred to as nutrient leaching and is a particular issue with nitrogen fertilisers (e.g. urea, potassium nitrate, ammonium nitrate) used to boost pasture plant and arable crop production (see Chapter 7). The rate of water flow through a soil (essentially the storage term in the water balance equation) is critical in controlling the rate of leaching and therefore has a major influence on water quality in a receiving stream.

SUMMARY

Water held in storage is an important part of the water balance equation. It is of particular importance as a change in storage, whether as an absorption term (negative) or a release (positive). As with all the processes in the hydrological cycle, storage is difficult to measure accurately at a useful spatial scale. This applies whether it is water held underground or as snow and ice. The release of water from storage may have a significant effect on river flows, as is demonstrated in the Mackenzie river Case Study (pp. 72–75).

ESSAY QUESTIONS

1 **Compare and contrast different methods for measuring soil water content at the hillslope scale.**
2 **Define the term *saturated hydraulic conductivity* and explain its importance in understanding groundwater flow.**
3 **Explain the terms *confined* and *unconfined* with respect to aquifers and describe how artesian wells come about.**
4 **Describe the field experiment required to assess the amount of snow in a seasonal snowpack and the timing of the snow melt.**

FURTHER READING

Ferguson, R.I. (1999) Snowmelt runoff models. *Progress in Physical Geography* **23**:205–228.
An overview of snow melt estimation techniques.

Freeze, R.A. and Cherry J.A. (1979) *Groundwater*. Prentice-Hall, Englewood Cliffs, NJ.
A classic text book on groundwater (including soil water).

Kendall, C. and McDonnell J.J. (eds) (1998) *Isotope tracers in catchment hydrology*. Elsevier Science, Amsterdam.
Detailed book on groundwater and streamwater ageing techniques.

Klute, A. (ed.) (1986) *Methods of soil analysis. Part 1: Physical and mineralogical methods*. American Society of Agronomy–Soil Science Society of America, Madison, Wisc.
A mine of information on soil methods.

Price, M. (1996) *Introducing groundwater* (2nd edn). Chapman and Hall, London.
An introductory text on groundwater.

5

RUNOFF

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the process of runoff leading to **channel flow**.
- A knowledge of the techniques for measuring streamflow and runoff directly.
- A knowledge of techniques used to estimate streamflow.

The amount of water within a river or stream is of great interest to hydrologists. It represents the end-product of all the other processes in the hydrological cycle and is where the largest amount of effort has gone into analysis of historical records. The methods of analysis are covered in Chapter 6; this chapter deals with the mechanisms that lead to water entering the stream: the runoff mechanisms. *Runoff* is a loose term that covers the movement of water to a channelised stream, after it has reached the ground as precipitation. The movement can occur either on or below the surface and at differing velocities. Once the water reaches a stream it moves towards the oceans in a channelised form, the process referred to as **streamflow** or **riverflow**. Streamflow is expressed as **discharge**: the volume of water over a defined time period. The SI units for discharge are m^3/s (*cumecs*). A continuous record of streamflow is called a **hydrograph** (see Figure 5.1). Although we think of this as continuous measurement it is normally either an averaged flow over a time period or a series of samples (e.g. hourly records).

In Figure 5.1 there are a series of peaks between periods of steady, much lower flows. The hydrograph peaks are referred to as **peakflow**, **stormflow** or even **quickflow**. They are the water in the stream during and immediately after a significant rainfall event. The steady periods between peaks are referred to as **baseflow** or sometimes slowflow (NB this is different from **low flow**; see Chapter 6). The shape of a hydrograph, and in particular the shape of the stormflow peak, is influenced by the storm characteristics (e.g. rainfall intensity and duration)

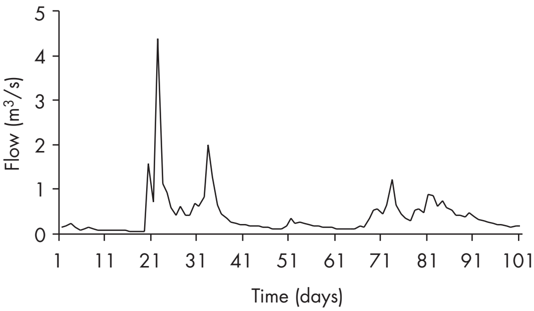

Figure 5.1 A typical hydrograph, taken from the river Wye, Wales for a 100-day period during the autumn of 1995. The values plotted against time are mean daily flow in cumecs.

and many physical characteristics of the upstream catchment. In terms of catchment characteristics the largest influence is exerted by catchment size, but other factors include slope angles, shape of catchment, soil type, vegetation type and percentage cover, degree of urbanisation and the antecedent soil moisture.

Figure 5.2 shows the shape of a storm hydrograph in detail. There are several important hydrological terms that can be seen in this diagram. The **rising limb** of the hydrograph is the initial steep part leading up to the highest or peakflow value. The water contributing to this part of the hydrograph is from *channel precipitation* (i.e. rain that falls directly onto the channel) and rapid runoff mechanisms. Some texts claim that channel precipitation shows up as a preliminary blip before the main rising limb.

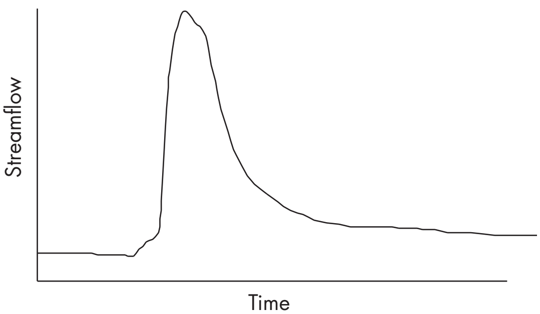

Figure 5.2 Demonstration storm hydrograph.

In reality this is very rarely observed, a factor of the complicated nature of storm runoff processes. The **recession limb** of the hydrograph is after the peak and is characterised by a long, slow decrease in streamflow until the baseflow is reached again. The recession limb is attenuated by two factors: storm water arriving at the mouth of a catchment from the furthest parts, and the arrival of water that has moved as underground flow at a slower rate than the streamflow.

Exactly how water moves from precipitation reaching the ground surface to channelised streamflow is one of the most intriguing hydrological questions, and one that cannot be answered easily. Much research effort in the past hundred years has gone into understanding runoff mechanisms; considerable advances have been made, but there are still many unanswered questions. The following section describes how it is believed runoff occurs, but there are many different scales at which these mechanisms are evident and they do not occur everywhere.

RUNOFF MECHANISMS

Figure 5.3 is an attempt to represent the different runoff processes that can be observed at the hillslope scale. **Overland flow** (Q_o) is the water which runs across the surface of the land before reaching the stream. In the subsurface, throughflow (Q_t) (some authors refer to this as **lateral flow**) occurs in the shallow subsurface, predominantly, although not always, in the unsaturated zone. Groundwater flow (Q_G) is in the deeper saturated zone. All of these are runoff mechanisms that contribute to streamflow. The relative importance of each is dependent on the catchment under study and the rainfall characteristics during a storm.

Overland flow

Some of the earliest research work on how overland flow occurs was undertaken by Robert Horton (1875–1945). In a classic paper from 1933, Horton

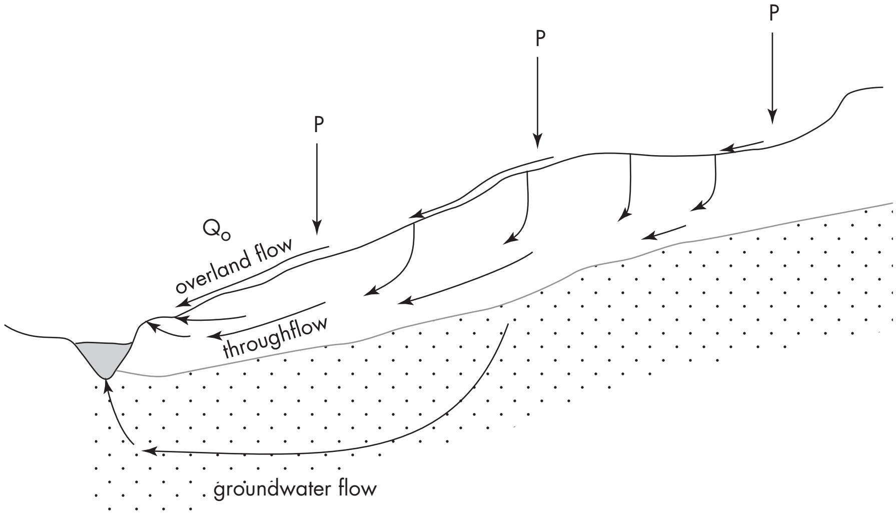

Figure 5.3 Hillslope runoff processes. See text for explanation of terms.

Source: Adapted from Dunne (1978)

hypothesised that overland flow occurred when the rainfall rate was higher than the infiltration rate of a soil. Horton went on to suggest that under these circumstances the excess rainfall collected on the surface before travelling towards the stream as a thin sheet of water moving across the surface. Under this hypothesis it is the infiltration rate of a soil that acts as a controlling barrier or partitioning device. Where the infiltration capacity of a soil is low, overland flow occurs readily. This type of overland flow is referred to as **infiltration excess overland flow** or **Hortonian overland flow** although as Beven (2004) points out, Horton himself referred to it as 'rainfall excess'.

Horton's ideas were extremely important in hydrology as they represented the first serious attempt to understand the processes of storm runoff that lead to a storm hydrograph. Unfortunately, when people started to measure infiltration capacities of soils they invariably found that they were far higher than most normal rainfall rates. This is illustrated in Table 5.1 where some typical infiltration capacities and rainfall rates are shown. Other measurements confirm high infiltration capacities for soils, e.g. Selby (1970) reports infiltration capacities of between 60 and 600 mm/hour on short grazed pasture on yellow-brown pumice soils in the central North Island of New Zealand. The values were higher for ungrazed grass and under trees and are generally higher than the measured rainfall intensities (Selby, 1970).

In addition to the infiltration capacity information, it is extremely rare to find a thin sheet of water moving over the surface during a storm event. It was observations such as those by Hursh (1944) and others that led to a general revision of Horton's hypothesis. Betson (1964) proposed the idea that within a catchment there are only limited areas that contribute overland flow to a storm hydrograph. This is referred to as the **partial areas concept**. Betson did not challenge the role of infiltration excess overland flow as the primary source of stormflow, but did challenge the idea of overland flow occurring as a thin sheet of water throughout a catchment.

Hewlett and Hibbert (1967) were the first to suggest that there might be another mechanism of overland flow occurring. This was particularly based

Table 5.1 Some typical infiltration rates compared to rainfall intensities

Soil and vegetation	*Infiltration rate (mm/hr)*	*Rainfall type*	*Rainfall intensity (mm/hr)*
Forested loam	100–200	Thunderstorm	50–100
Loam pasture	10–70	Heavy rain	5–20
Sand	3–15	Moderate rain	0.5–5
Bare clay	0–4	Light rain	0.5

Source: From Burt (1987)

on the observations from the eastern USA: that during a storm it was common to find all the rainfall infiltrating a soil. Hewlett and Hibbert (1967) hypothesised that during a rainfall event all the water infiltrated the surface. This hypothesis was confirmed by a comprehensive field study by Dunne and Black (1970).

Through a mixture of infiltration and throughflow the water table would rise until in some places it reached the surface. At this stage overland flow occurs as a mixture of return flow (i.e. water that has been beneath the ground but returns to the surface) and rainfall falling on saturated areas. This type of overland flow is referred to as **saturated overland flow**. Hewlett and Hibbert (1967) suggested that the water table was closest to the surface, and therefore likely to rise to the surface quickest, adjacent to stream channels and at the base of slopes. Their ideas on stormflow were that the areas contributing water to the hydrograph peaks were the saturated zones, and that these vary from storm to storm. In effect the saturated areas immediately adjacent to the stream act as extended channel networks. This is referred to as the **variable source areas concept**. This goes a step beyond the ideas of Betson (1964) as the catchment has a partial areas response but the response area is dynamic; i.e. variable in space and time.

So who was right: Horton, or Hewlett and Hibbert? The answer is that both were. Table 5.2 provides a summary of the ideas for storm runoff generation described here. It is now accepted that saturated overland flow (Hewlett and Hibbert) is the dominant overland flow mechanism in humid, mid-latitude areas. It is also accepted that the variable source areas concept is the most valid description of stormflow processes. However, where the infiltration capacity of a soil is low or the rainfall rates are high, Hortonian overland flow does occur. In Table 5.1 it can be seen that there are times when rainfall intensities will exceed infiltration rates under natural circumstances. In arid and semi-arid zones it is not uncommon to find extremely high rainfall rates (fed by convective storms) that can lead to infiltration excess overland flow and rapid flood events; this is called **flash flooding**.

Table 5.2 A summary of the ideas on how stormflow is generated in a catchment

	Horton	*Betson*	*Hewlett and Hibbert*
Infiltration	Controls overland flow	Controls overland flow	All rainfall infiltrates
Overland flow mechanism	Infiltration excess	Infiltration excess	Saturated overland flow
Contributing area	Uniform throughout the catchment	Restricted to certain areas of the catchment	Contributing area is variable in time and space

Examples of low infiltration rates can be found with compacted soils (e.g. from vehicle movements in an agricultural field), on roads and paved areas, on heavily crusted soils and what are referred to as **hydrophobic soils**.

Basher and Ross (2001) reported infiltration capacities of 400 mm/hour in market gardens in the North Island of New Zealand and that these rates increased during the growing season to as high as 900 mm/hour. However, Basher and Ross (2001) also showed a decline in infiltration capacity to as low as 0.5 mm/hour in wheel tracks at the same site.

Hydrophobic soils have a peculiar ability to swell rapidly on contact with water, which can create an impermeable barrier at the soil surface to infiltrating water, leading to Hortonian overland flow. The cause of hydrophobicity in soils has been linked into several factors including the presence of micro-rhizal fungi and swelling clays such as allophane (Doerr *et al*., 2007). Hydrophobicity is a temporary soil property; continued contact with water will increase the infiltration rate. For example Clothier *et al*. (2000) showed how a yellow brown earth/loam changed from an initial infiltration capacity of 2 mm/hour to 14 mm/hour as the soil hydrophobicity breaks down.

In Hewlett and Hibbert's (1967) original hypothesis it was suggested that contributing saturated areas would be immediately adjacent to stream channels. Subsequent work by the likes of Dunne and Black (1970), Anderson and Burt (1978) and others has identified other areas in a catchment prone to inducing saturated overland flow. These include hillslope hollows, slope concavities (in section) and where there is a thinning of the soil overlying an impermeable base. In these situations any throughflow is likely to return to the surface as the volume of soil receiving it is not large enough for the amount of water entering it. This can be commonly observed in the field where wet and boggy areas can be found at the base of slopes and at valley heads (hillslope hollows).

Subsurface flow

Under the variable source areas concept there are places within a catchment that contribute overland flow to the storm hydrograph. When we total up the amount of water found in a storm hydrograph it is difficult to believe that it has all come from overland flow, especially when this is confined to a relatively small part of the catchment (i.e. variable source areas concept). The manner in which the recession limb of a hydrograph attenuates the stormflow suggests that it may be derived from a slower movement of water: subsurface flow. In addition to this, tracer studies looking at where the water has been before entering the stream as stormflow have found that a large amount of the storm hydrograph consists of 'old water' (e.g. Martinec *et al*., 1974; Fritz *et al*., 1976). This old water has been sitting in the soil, or as fully saturated groundwater, for a considerable length of time and yet enters the stream during a storm event. There have been several theories put forward to try and explain these findings, almost all involving throughflow and groundwater.

Throughflow is a general term used to describe the movement of water through the unsaturated zone; normally this is the soil matrix. Once water infiltrates the soil surface it continues to move, either through the soil matrix or along preferential flow paths (referred to as lateral or preferential flow). The rate of soil water movement through a saturated soil matrix is described by Darcy's law (see Chapter 4) and the Richards approximation of Darcy's law when below saturation. Under normal, vertical, infiltration conditions the hydraulic gradient has a value of –1 and the saturated hydraulic conductivity is the infiltration capacity. Once the soil is saturated the movement of water is not only vertical. With a sloping water table on a hillsope, water moves down slope. However, the movement of water through a saturated soil matrix is not rapid, e.g. Kelliher and Scotter (1992) report a K_{sat} value of 13 mm/hour for a fine sandy loam. In order for throughflow to contribute to storm runoff there must be another mechanism (other than matrix flow) operating.

One of the first theories put forward concerning the contribution of throughflow to a storm hydrograph was by Horton and Hawkins (1965) (this Horton was a different person from the proposer of Hortonian overland flow). They proposed the mechanism of *translatory* or *piston flow* to explain the rapid movement of water from the subsurface to the stream. They suggested that as water enters the top of a soil column it displaces the water at the bottom of the column (i.e. old water), and the displaced water enters the stream. The analogy is ~~dra~~wn to a piston where pressure at the top of the piston chamber leads to a release of pressure at the bottom. The release of water to the stream can be modelled as a pressure wave rather than tracking individual particles of water. Piston flow has been observed in laboratory experiments with soil columns (e.g. Germann and Beven, 1981).

At first glance the simple piston analogy seems unlike a real-life situation since a hillslope is not bounded by impermeable sides in the same way as a piston chamber. However the theory is not as far-fetched as it may seem, as the addition of rainfall infiltrating across a complete hillslope is analogous to pressure being applied from above and in this case the boundaries are upslope (i.e. gravity) and the bedrock below. Brammer and McDonnell (1996) suggest that this may be a mechanism for the rapid movement of water along the bedrock and soil interface on the steep catchment of Maimai in New Zealand. In this case it is the hydraulic gradient created by an addition of water to the bottom of the soil column, already close to saturated, that forces water along the base where hydraulic conductivities are higher.

Ward (1984) draws the analogy of a thatched roof to describe the contribution of subsurface flow to a stream (based on the ideas of Zaslavsky and Sinai, 1981). When straw is placed on a sloping roof it is very efficient at moving water to the bottom of the roof (the guttering being analogous to a stream) without visible overland flow. This is due to the preferential flow direction along, rather than between, sloping straws. Measurements of hillslope soil properties do show a higher hydraulic conductivity in the downslope rather than vertical direction. This would account for a movement of water downslope as throughflow, but it is still bound up in the soil matrix and reasonably slow.

There is considerable debate on the role of **macropores** in the rapid movement of water through the soil matrix. Macropores are larger pores within a soil matrix, typically with a diameter greater than 3 mm. They may be caused by soils cracking, worms burrowing or other biotic activities. The main interest in them from a hydrologic point of view is that they provide a rapid conduit for the movement of water through a soil. The main area of contention concerning macropores is whether they form continuous networks allowing rapid movement of water down a slope or not. There have been studies suggesting macropores as a major mechanism contributing water to stormflow (e.g. Mosley, 1979, 1982; Wilson *et al*., 1990), but it is difficult to detect whether these are from small areas on a hillslope or continuous throughout. Jones (1981) and Tanaka (1992) summarise the role of pipe networks (a form of continuous macropores) in hillslope hydrology. Where found, pipe networks have considerable effect on the subsurface hydrology but they are not a common occurrence in the field situation.

The role of macropores in runoff generation is unclear. Although they are capable of allowing rapid movement of water towards a stream channel there is little evidence of networks of macropores moving large quantities of water in a continuous fashion. Where macropores are known to have a significant role is in the rapid movement of water to the saturated layer (e.g. Heppell *et al*., 1999) which may in turn lead to piston flow (McGlynn *et al*., 2002).

Groundwater contribution to stormflow

Another possible explanation for the presence of old water in a storm hydrograph is that it comes from the saturated zone (groundwater) rather than from throughflow. This is contrary to conventional hydrological wisdom which suggests that groundwater contributes to baseflow but not to the stormflow component of a hydrograph. Although a groundwater contribution to stormflow had been suggested before, it was not until Sklash and Farvolden (1979) provided a theoretical mechanism for this to occur that the idea was seriously considered. They proposed the capillary fringe hypothesis to explain the groundwater ridge, a rise in the water table immediately adjacent to a stream (as observed by Ragan, 1968). Sklash and Farvolden (1979) suggested that the addition of a small amount of infiltrating rainfall to the zone immediately adjacent to a stream causes the soil water to move from an unsaturated state (i.e. under tension) to a saturated state (i.e. a positive pore pressure expelling water). As explained in Chapter 4, the relationship between soil water content and soil water tension is non-linear. The addition of a small amount of water can cause a rapid change in soil moisture status from unsaturated to saturated. This provides the groundwater ridge which:

> not only provides the early increased impetus for the displacement of the groundwater already in a discharge position, but it also results in an increase in the size of the groundwater discharge area which is essential in producing large groundwater contributions to the stream.
>
> (Sklash and Farvolden, 1979: 65)

An important point to stress from the capillary fringe hypothesis is that the groundwater ridge is developing well before any throughflow may have been received from the contributing hillslope areas. These ideas confirm the variable source areas concept and provide a mechanism for a significant old water contribution to storm hydrographs. Field studies such as that by McDonnell (1990) have observed groundwater ridging to a limited extent, although it is not an easy task as often the instrument response time is too slow to detect the rapid change in pore pressure properly.

Case study

THE MAIMAI RUNOFF GENERATION STUDIES

Figure 5.4 Maimai catchments in South Island of New Zealand. At the time of photograph (1970s) five catchments had been logged and are about to be replanted with *Pinus radiata*.

The Maimai catchment study (near Reefton on West Coast of the South Island of New Zealand) was established in 1974 for research into the effects of logging native beech forest (*Nothofagus*) and replanting with different non-indigenous species (Figure 5.4). The installation of hydrological measuring equipment and the fact that rainfall and stormflow are frequently observed made it an ideal place for studying stormflow generation mechanisms in depth. The knowledge gained from detailed hydrological process studies at Maimai have played a major part in shaping thinking on stormflow generation mechanisms.

The Maimai catchment is characterised by short, steep slopes (approximately 300 m with angles of around 35°), covered in thick vegetation, with incised channels and very small valley bottoms. Annual rainfall is approximately 2,600 mm with an average of 156 rain days a year, and stormflow makes up 65 per cent of the total streamflow (Rowe *et al*., 1994; Pearce *et al*., 1986).

Mosley (1979, 1982) used Maimai to investigate the role of macropores as conduits for rapid movement of rainfall to the stream. Observations of macropore flow rates using cut soil faces and dye tracers suggested that rainfall could travel down the short steep hillslopes at Maimai in less than 3 hours (i.e. within the time frame of a storm event). Subsequent chemical and isotopic analysis of streamflow, rainfall and water exiting the cut soil pit faces showed that the majority of measured streamflow was 'old' water, suggesting that rapid, extensive macropore flow was not the main mechanism for stormflow generation (Pearce *et al*., 1986).

McDonnell (1990) investigated this further, in particular looking at possible groundwater ridging (Sklash and Farvolden, 1979) as a mechanism for large amounts of old water as saturated overland flow. Although this could be observed at Maimai, the amount of water held near the stream prior to an event was not large enough to account for all of the old water, which suggested that another mechanism (e.g. piston flow) might be working (McDonnell, 1990).

McGlynn *et al*. (2002) present a summary conceptual diagram of runoff mechanisms on Maimai hillslopes that combines many of the features described above (see Figure 5.5). In this model there is rapid infiltration of water through macropores to reach the bedrock. At this stage a form of piston flow occurs as the saturated zone at the base of the soil mantle is confined by the soil matrix above it. At the bedrock interface there may be a network of macropores or else the same situation of a confined aquifer in that the soil matrix above has a much lower hydraulic conductivity. Water is then pushed out at the bottom due to the pressure from new water arriving directly at the bedrock interface. There is also a mixing of the new water with old water sitting in bedrock hollows, creating a rapid movement of old water into the stream during storm events.

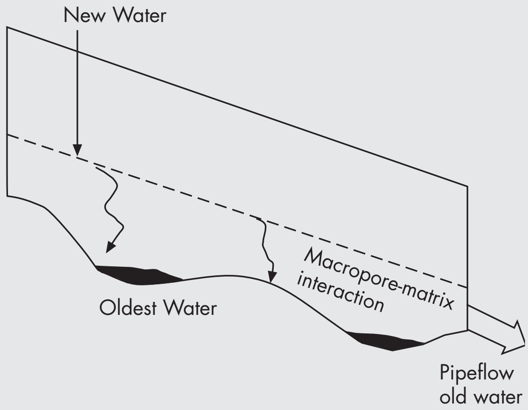

Figure 5.5 Summary hypothesis for hillslope stormflow mechanisms at Maimai. Rapid movement of water occurs through rapid infiltration to the bedrock interface and then a form of piston flow along this interface.

Source: adapted from McGlynn *et al*. (2002)

How relevant are the Maimai stormflow generation studies?

The studies that have taken place at Maimai have been extremely important in influencing hydrological thinking around the world. However, an argument can be made that the conditions at Maimai are far from generally applicable elsewhere. The main study catchment (M8) has short, steep slopes and is in an area of high, and frequent, rainfall. The soils are extremely porous (infiltration rates in excess of 1,600 mm/hour have been measured) and remain within 10 per cent of saturation for most of the year (Mosley, 1979). These conditions are not common and it would be difficult to generalise the concepts beyond Maimai. One of the really important concepts that Maimai has shown is that under conditions ideal for stormflow generation the mechanisms are still extremely complex and spatially variable. This is true wherever in the world the study is taking place.

Summary of storm runoff mechanisms

The mechanisms that lead to a storm hydrograph are extremely complex and still not fully understood. Although this would appear to be a major failing in a science that is concerned with the movement of water over and beneath the surface, it is also an acknowledgement of the extreme diversity found in nature. In general there is a reasonable understanding of possible storm runoff mechanisms but it is not possible to apply this universally. In some field situations the role of throughflow and piston flow are important, in others not; likewise for groundwater contributions, overland flow and pipeflow. The challenge for modern hydrology is to identify quickly the dominant mechanisms for a particular hillslope or catchment so that the understanding of the hydrological processes in that situation can be used to aid management of the catchment.

The processes of storm runoff generation described here are mostly observable at the hillslope scale. At the catchment scale (and particularly for large river basins) the timing of peak flow (and consequently the shape of the storm hydrograph) is influenced more by the channel drainage network and the precipitation characteristics of a storm than by the mechanisms of runoff. This is a good example of the problem of scale described in Chapter 1. At the small hillslope scale storm runoff generation mechanisms are important, but they become considerably less so at the much larger catchment scale.

Baseflow

In sharp contrast to the storm runoff debate, there is general consensus that the major source of baseflow is groundwater – and to a lesser extent throughflow. This is water that has infiltrated the soil surface and moved towards the saturated zone. Once in the saturated zone it moves downslope, often towards a stream. A stream or lake is often thought to occur where the regional water table intersects the surface, although this may not always be the case. In Chapter 4 the relationship between groundwater and streamflow has been explained (see Figure 4.9). However in general it can be said that baseflow is provided by the slow seepage of water from groundwater into streams. This will not necessarily be visible (e.g. springs) but can occur over a length of streambank and bed and is only detectable through repeated measurement of streamflow down a reach.

Channel flow

Once water reaches the stream it will flow through a channel network to the main river. The controls over the rate of flow of water in a channel are to do with the volume of water present, the gradient of the channel, and the resistance to flow experienced at the channel bed. This relationship is described in uniform flow formulae such as the Chezy and Manning equations (see p. 92). The resistance to flow is governed by the character of the bed surface. Boulders and vegetation will create a large amount of friction, slowing the water down as it passes over the bed.

In many areas of the world the channel network is highly variable in time and space. Small channels may be ephemeral and in arid regions will frequently only flow during flood events. The resistance to flow under these circumstances is complicated by the infiltration that will be occurring at the water front and bed surface. The first flush of water will infiltrate at a much higher rate as it fills the available pore space in the soil/rock at the bed surface. This will remove water from the stream and also slow the water front down as it creates a greater friction surface. Under a continual flow regime the infiltration from the stream to ground will depend on the hydraulic gradient and the infiltration capacity.

MEASURING STREAMFLOW

The techniques and research into the measurement of streamflow are referred to as **hydrometry**. Streamflow measurement can be subdivided into two important subsections: instantaneous and continuous techniques.

Instantaneous streamflow measurement

Velocity–area method

Streamflow or discharge is a volume of water per unit of time. The standard units for measurement of discharge are m^3/s (cubic metres per second or *cumecs*). If we rewrite the units of discharge we can think of them as a water velocity (m/s) passing through a cross-sectional area (m^2). Therefore:

$$m^3/s = m/s \times m^2 \tag{5.1}$$

The **velocity–area method** measures the stream velocity, the stream cross-sectional area and multiplies the two together. In practice this is carried out by dividing the stream into small sections and measuring the velocity of flow going through each cross-sectional area and applying equation 5.2.

$$Q = v_1a_1 + v_2a_2 + \ldots v_ia_i \tag{5.2}$$

where Q is the streamflow or discharge (m^3/s), v is the velocity measured in each trapezoidal cross-sectional area (see Figure 5.6), and a is the area of the trapezoid (usually estimated as the average of two depths divided by the width between).

The number of cross-sectional areas that are used in a discharge measurement depend upon the width and smoothness of stream bed. If the bed is particularly rough it is necessary to use more cross-sectional areas so that the estimates are as close to reality as possible (note the discrepancy between the broken and solid lines in Figure 5.6).

The water velocity measurement is usually taken with a flow meter (Figure 5.7). This is a form of propeller inserted into the stream which records the number of propeller turns with time. This reading can be easily converted into a stream velocity using the calibration equation supplied with the flow meter.

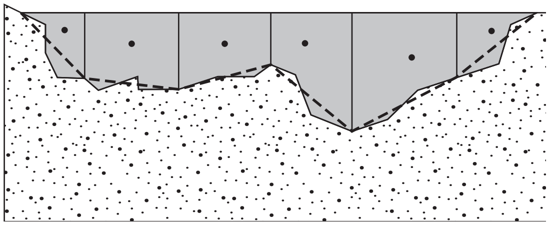

Figure 5.6 The velocity–area method of streamflow measurement. The black circles indicate the position of velocity readings. Dashed lines represent the triangular or trapezoidal cross-sectional area through which the velocity is measured.

Figure 5.7 Flow gauging a small stream.

In the velocity–area method it is necessary to assume that the velocity measurement is representative of all the velocities throughout the cross-sectional area. It is not normally possible to take multiple measurements so an allowance has to be made for the fact that the water travels faster along the surface than nearer the stream bed. This difference in velocity is due to friction exerted on the water as it passes over the stream bed, slowing it down. As a general rule of thumb the sampling depth should be 60 per cent of the stream depth – that is, in a stream that is 1 m deep the sampling point should be 0.6 m below the surface or 0.4 m above the bed. In a deep river it is good practice to take two measurements (one at 20 per cent and the other at 80 per cent of depth) and average the two.

Where there is no velocity meter available it may be possible to make a very rough estimate of stream velocity using a float in the stream (i.e. the time it takes to cover a measured distance). When using this method allowance must be made for the fact that the float is travelling on the surface of the stream at a faster rate than water closer to the stream bed.

The velocity–area method is an effective technique for measuring streamflow in small rivers, but its reliability is heavily dependent on the sampling strategy. The technique is also less reliable in small, turbid streams with a rough bed (e.g. mountain streams). Under these circumstances other streamflow estimation techniques such as **dilution gauging** may be more applicable (see streamflow estimation section).

Continuous streamflow measurement

The methods of instantaneous streamflow measurement described above only allow a single measurement to be taken at a location. Although this can be repeated at a future date it requires a continuous measurement technique to give the data for a hydrograph. There are three different techniques that can be used for this method: stage discharge relationships, flumes and weirs, and ultrasonic flow gauging.

Stage vs discharge relationship

River **stage** is another term for the water level or height. Where multiple discharge measurements have been taken (i.e. repeat measurements using velocity–area method) it is possible to draw a relationship between river stage and discharge: the so-called **rating curve**. An example of a rating curve is shown in Figure 5.8. This has the advantage of allowing continuous measurement of river stage (a relatively simple task) that can then be equated to the actual discharge. The stage discharge relationship is derived through a series of velocity–area measurements at a particular site while at the same time recording the stage with a stilling well (see Figure 5.9). As can be seen in Figure 5.8, the rating curve is non-linear, a reflection of the river bank profile. As the river fills up between banks it takes a greater volume of water to cause a change in stage than at low levels.

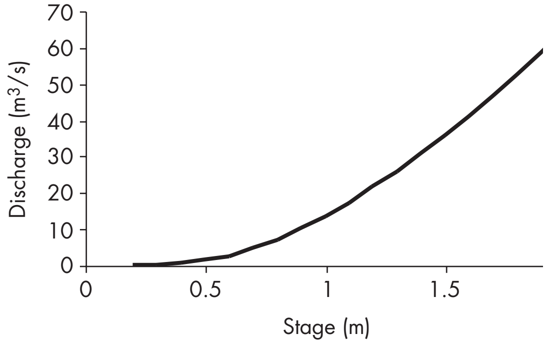

Figure 5.8 A rating curve for the river North Esk in Scotland based on stage (height) and discharge measurements from 1963–1990.

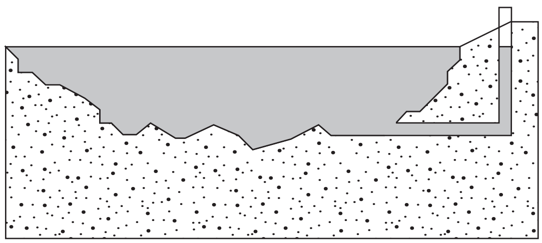

Figure 5.9 Stilling well to provide a continuous measurement of river stage (height). The height of water is measured in the well immediately adjacent to the river.

An accurate stage vs discharge relationship is dependent on frequent and accurate measurement of river discharge, and a static river bed profile. If the river bed profile changes (e.g. during a large flood event it may get scoured out or new sediment deposited), the stage vs discharge relationship will change and the historic relationship will no longer be valid. This assumption of a static river bed profile can sometimes be problematic, leading to the installation of a concrete structure (e.g. flume or weir) to maintain stability.

One of the difficulties with the stage vs discharge relationship is that the requirement of frequent measurements of river discharge lead to many measurements taken during periods of low and medium flow but very few during flood events. This is for the double reason that: floods are infrequent and unlikely to be measured under a regular monitoring programme; and the danger of streamflow gauging during a flood event. The lack of data at the extreme end of the stage vs discharge curve may lead to difficulties in interpreting data during peak flows. The error involved in estimating peak discharge from a measured stage vs discharge relationship will be much higher at the high flow end of curve.

When interpreting data derived from the stage discharge vs relationship it is important that the hydrologist bears in mind that it is stream stage that is being measured and from this stream discharge is inferred (i.e. it is not a direct measurement of stream discharge).

Flumes and weirs

Flumes and weirs utilise the stage–discharge relationship described above but go a step further towards providing a continuous record of river discharge. If we think of stream discharge as consisting of a river velocity flowing through a cross-sectional area (as in the velocity profile method) then it is possible to isolate both of these terms separately. This is what flumes and weirs, or *stream gauging structures*, attempt to do.

The first part to isolate is the stream velocity. The way to do this is to slow a stream down (or, in some rare cases, speed a stream up) so that it flows with constant velocity through a known cross-sectional area. The critical point is that in designing a flume or weir the river flows at the same velocity (or at least a known velocity) through the gauging structure irrespective of how high the river level is. Although this seems counter-intuitive (rivers normally flow faster during flood events) it is achievable if there is an area prior to the gauging structure that slows the river down: a stilling pond.

The second part of using a gauging structure is to isolate a cross-sectional area. To achieve this a rigid structure is imposed upon the stream so that it always flows through a known cross-sectional area. In this way a simple measure of stream height through the gauging structure will give the cross-sectional area. Stream height is normally derived through a stilling well, as described in Figure 5.9, except in this case there is a regular cross-sectional area.

Once the velocity and cross-sectional area are kept fixed the rating curve can be derived through a mixture of experiment and hydraulic theory. These relationships are normally power equations dependent on the shape of cross-sectional area used in the flume or weir. There is an international standard for manufacture and maintenance of weirs (ISO 1438) that sets out theoretical ratings curves for different types of structures. The general formula for a V notch weir is shown in equation 5.3.

$$Q = 0.53.\sqrt{2g}.C.\tan\left(\frac{\theta}{2}\right).h^{2.5} \tag{5.3}$$

where Q is discharge (m^3/s); g is the acceleration due to gravity (9.81 m/s^2); C is coefficient of discharge (see Figure 5.10); θ is the angle of V-notch (°); h is the height of water or stage (m). The coefficient of discharge can be estimated from figure 5.10 for a certain angle of V-notch. For a 90° V-notch the coefficient of discharge is 0.578 and the rating equation becomes:

$$Q = 1.366h^{2.5} \tag{5.4}$$

There is a similar type of equation for rectangular weirs, based on the width of the rectangular exit and another version of the coefficient of discharge relationship.

The shape of cross-sectional area is an important consideration in the design of flumes and weirs. The shape of permanent structure that the river flows

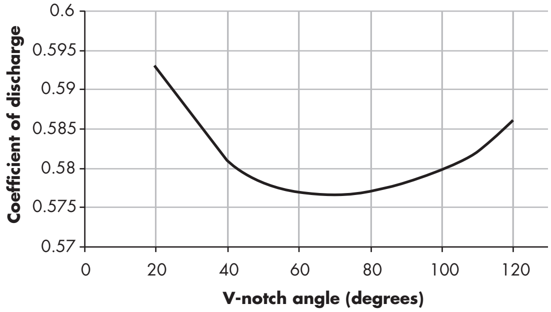

Figure 5.10 Coefficient of discharge for V-notch weirs (ISO 1438).

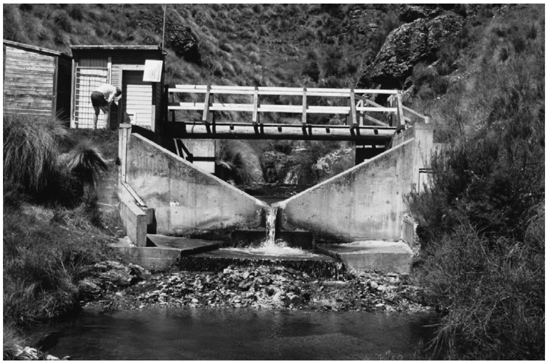

Figure 5.11 A V-notch weir. The water level in the pond behind the weir is recorded continuously.

Figure 5.12 A trapezoidal flume. The stream passes through the flume and the water level at the base of the flume is recorded continuously.

through is determined by the flow regime of the river and the requirements for the streamflow data. A common shape used is based on the V-structure (see Figure 5.11). The reason for this is that when river levels are low, a small change in river flow will correspond to a significant change in stage (measured in the stilling well). This sensitivity to low flows makes data from this type of flume or weir particularly suitable for studying low flow hydrology. It is important that under high flow conditions the river does not overtop the flume or weir structure. The V shape is convenient for this also because as discharge increases the cross-sectional area flowed through increases in a non-linear fashion. The angle of the V-notch will vary depending on the size of stream being measured and the sensitivity required (90° and 120° V-notch weirs are both commonly used).

One of the difficulties in maintaining gauging structures is that by slowing the river down in a stilling pond any sediment being carried by the water may be deposited (see **Hjulstrom curve** in Chapter 7), which in time will fill the stilling pond and lessen its usefulness. Because of this the stilling pond needs to be dredged regularly, particularly in a high energy environment such as mountain streams. To overcome this difficulty there is a design of trapezoidal flume that speeds the stream up rather than slows it down (see Figure 5.12). The stream is forced to go down a steep section immediately prior to the gauging structure. In this way any sediment is flushed out of the weir, removing the need for regular dredging. This is really only possible for small streams as the power of large rivers at high velocities would place enormous strains on the gauging structure.

The difference between flumes and weirs

Although flumes and weirs perform the same function – measuring stream discharge in a continuous fashion – they are not the same. In a weir the water is forced to drop over a structure (the weir – Figure 5.11) in the fashion of a small waterfall. In a flume (Figure 5.12) the water passes through the structure without having a waterfall at the end.

Ultrasonic flow gauging

Recent technological developments have led to the introduction of a method of measuring stream discharge using the properties of sound wave propagation in water. The method actually measures water velocity, but where the stream bed cross-sectional area is known (and constant) the instrumentation can be left in place and combined with measurements of stage to provide a continuous measurement of river discharge. There are two types of **ultrasonic flow gauges** that work in slightly different ways.

The first method measures the time taken for an ultrasonic wave emitted from a transmitter to reach a receiver on the other side of a river. The faster the water speed, the greater the deflection of the wave path and the longer it will take to cross the river. Sound travels at approximately 1,500 m/s in water (dependent on water purity and depth) so the instrumentation used in this type of flow gauging needs to be extremely precise and be able to measure in nanoseconds. This type of flow gauging can be installed as a permanent device but needs a width of river greater than 5 m and becomes unreliable with a high level of suspended solids.

The second method utilises the Doppler effect to measure the speed of particles being carried by the stream. At an extremely simple level this is a measurement of the wavelength of ultrasonic waves that bounce off suspended particles – the faster the particle the shorter the wavelength. This type of instrument works in small streams (less than 5 m width) and requires some suspended matter.

Measuring hillslope runoff

The measurement of runoff may be required to assess the relative contribution of different hillslope runoff processes; i.e. throughflow, overland flow, etc. There are no standard methods for the measurement of runoff processes; different researchers use different techniques according to the field conditions expected and personal preference.

Overland flow

The amount of water flowing over the soil surface can be measured using collection troughs at the bottom of hillslopes or runoff plots. A runoff plot is an area of hillslope with definite upslope and side boundaries so that you can be sure all the overland flow is generated from within each plot. The upslope and side boundaries can be constructed by driving metal plates into the soil and leaving them protruding above the surface. It is normal to use several runoff plots to characterise overland flow on a slope as it varies considerably in time and space. This spatial and temporal variation may be overcome with the use of a rainfall simulator.

Throughflow

Measurement of throughflow is fraught with difficulty. The only way to measure it is with throughflow troughs dug into the soil at the appropriate height. The problem with this is that in digging, the soil profile is disturbed and consequently the flow characteristics change. It is usual to insert troughs into a soil face that has been excavated and then refill the hole. This may still overestimate throughflow as the reconstituted soil in front of the troughs may encourage flow towards it as an area allowing rapid flow.

ESTIMATING STREAMFLOW

In the past thirty years probably the greatest effort in hydrological research has gone into creating numerical models to simulate streamflow. With time these have developed into models simulating all the processes in the hydrological cycle so that far more than just streamflow can be estimated. However, it is often streamflow that is seen as the end-product of a model, a reflection of the importance streamflow has as a hydrological parameter. These models are described in Chapter 6; this section concentrates on direct estimates of streamflow.

Physical or geomorphological estimation techniques

The geomorphological approach to river systems utilises the idea that the river channel is in equilibrium with the flow regime. This suggests that measures of the channel (e.g. depth/width ratio, **wetted perimeter**, height to **bankfull discharge**) can be used to estimate the streamflows in both a historical and contemporary sense. Wharton (1995) provides a review of these different techniques. This is not a method that can be used to estimate the discharge in a river at one particular time, but it can be used to estimate parameters such as the mean annual **flood**. Important parameters to consider are the stream diameter, wetted perimeter and average depth. This is particularly for the area of channel that fills up during a small flooding event: so-called bankfull discharge.

It is possible to estimate the average velocity of a river stretch using a kinematic wave equation such as Manning's (equation 5.5).

$$v = \frac{k.R^{2/3}.\sqrt{s}}{n} \qquad (5.5)$$

where v is velocity (m/s); k is a constant depending on which units of measurement are being used (1 for SI units, 1.49 for Imperial); R is the **hydraulic radius** (m); s is the slope (m/m); and n is the Manning roughness coefficient. Hydraulic radius is the wetted perimeter of a river divided by the cross-sectional area. In very wide channels this can be approximated as mean depth (Goudie *et al*., 1994). The Manning roughness coefficient is estimated from knowledge of the channel characteristics (e.g. vegetation and bed characteristics) in a similar manner to Chezy's roughness coefficient in Table 5.3. Tables of Manning roughness coefficient can be found in Richards (1982), Maidment (1992), Goudie *et al*. (1994), and in other fluvial geomorphological texts.

Dilution gauging

Dilution gauging works on the principle that the more water there is in a river the more it will dilute a solute added into the river. There is a well-

Table 5.3 Chezy roughness coefficients for some typical streams

Type of channel	*Chezy roughness coefficient for a hydraulic radius of 1 m*
Artificial concrete channel	71
Excavated gravel channel	40
Clean regular natural channel <30 m wide	33
Natural channel with some weeds or stones <30 m wide	29
Natural channel with sluggish weedy pools <30 m wide	14
Mountain streams with boulders	20
Streams larger than 30 m wide	40

Source: Adapted from Richards (1982)

established relationship between the amount of the tracer found naturally in the stream (C_o), the concentration of tracer put into the river (C_t), the concentration of tracer measured downstream after mixing (C_d), and the stream discharge (Q). The type of tracer used is dependent on the equipment available; the main point is that it must be detectable in solution and non-harmful to the aquatic flora and fauna. A simple tracer that is often used is a solution of table salt (NaCl), a conductivity meter being employed to detect the salt solution.

There are two different ways of carrying out dilution gauging that use slightly different equations. The first puts a known volume of tracer into the river and measures the concentration of the 'slug' of tracer as it passes by the measurement point. This is referred to as gulp dilution gauging. The equation for calculating flow by this method is shown in equation 5.6.

$$Q = \frac{C_t V}{\sum\left(C_d - C_o\right)\Delta t} \tag{5.6}$$

where Q is the unknown streamflow, C is the concentration of tracer either in the slug (t), downstream (d), or background in the stream (o); Δt is the time interval. The denominator of this equation is the sum of measured concentrations of tracer downstream.

The second method uses a continuous injection of tracer into the river and measures the concentration downstream. The continuous injection method is better than the slug injection method as it measures the concentration over a greater length of time, however it requires a large volume of the tracer. Using the formula listed below the stream discharge can be calculated using equation 5.7.

$$Q = q\frac{C_t - C_d}{C_d - C_0} \tag{5.7}$$

where q is the flow rate of the injected tracer (i.e. injection rate) and all other terms are as for the gulp injection method.

Probably the most difficult part of dilution gauging is calculating the distance downstream between where the tracer is injected and the river concentration measuring point (the mixing distance). This can be estimated using equation 5.8.

$$L = 0.13C_z\left(\frac{0.7C_z + w}{g}\right)\left(\frac{w^2}{d}\right) \tag{5.8}$$

where L = mixing distance (m)
C_Z = Chezy's **roughness coefficient** (see Table 5.3)
w = average stream width (m)
g = gravity constant ($\approx$9.8 m/s^2)
d = average depth of flow (m)

FLOODS

The term *flood* is difficult to define except in the most general of terms. In a river a flood is normally considered to be an inundation of land adjacent to a river caused by a period of abnormally large discharge or encroachment by the sea (see cover photograph, Figure 5.13, and Plate 6), but even this definition is fraught with inaccuracy. Flooding may occur from sources other than rivers (e.g. the sea and lakes), and 'abnormal' is difficult to pin down, particularly within a timeframe. Floods come to our attention through the amount of damage that they cause and for this reason they are often rated on a cost basis rather than on hydrological criteria. Hydrological and monetary assessments of flooding often differ markedly because the economic valuation is highly dependent on location. If the area of land inundated by a flooding river is in an expensive region with large infrastructure then the cost will be considerably higher than, say, for agricultural land. Two examples of large-scale floods during the 1990s illustrate this point. In 1998 floods in China caused an estimated US$20 billion of damage with over 15 million people being displaced and 3,000 lives lost (Smith, 2001). This flood was on a similar scale to one that occurred in the same region during 1954. A much larger flood (in a hydrological sense)

Figure 5.13 Images of flood inundation in Fiji, 2007.

in the Mississippi and Missouri rivers during 1993 resulted in a similar economic valuation of loss (US$15–20 billion) but only 48 lives were lost (USCE, 1996). The flood was the highest in the hydrological record and had an average recurrence interval of between 100 and 500 years (USCE, 1996). The difference in lives lost and relative economic loss (for size of flood) is a reflection of the differing response to the flood in two economically contrasting countries.

As described in Chapter 2 for precipitation, flooding is another example where the *frequency–magnitude relationship* is important. Small flood events happen relatively frequently whereas the really large floods occur rarely but cause the most damage. The methods for interpreting river flows that may be used for flood assessment are discussed in Chapter 6. They provide some form of objective flood size assessment, but their value is highly dependent on the amount of data available.

Floods are a frequently occurring event around the world. At the time of preparing of this chapter (June and July 2007) there were eleven large flood events reported in the news media (see Table 5.4). These floods were caused by varying amounts of rainfall, and occurred in different seasons of the year but all caused significant damage and in many cases loss of lives. There are numerous reasons why a river will flood and they almost always relate back to the processes found within the hydrological cycle. The main cause of river floods is when there is too much rainfall for the river to cope with. Other, more special causes of floods are individual events like dam bursts, **jökulhlaups** (ice-dam bursts) or snow melt (see pp. 72–75).

Influences on flood size

The extent and size of the flood can often be related to other contributing factors that increase the effect of high rainfall. Some of these factors are described here but all relate back to concepts introduced in earlier chapters detailing the processes found within the hydrological cycle. Flooding provides an excellent example of the importance of scale, introduced in Chapter 1. Many of the factors discussed here have an influence at the small scale (e.g. hillslopes or small research catchments of less than 10 km^2) but not at the larger overall river catchment scale.

Antecedent soil moisture

The largest influence on the size of a flood, apart from the amount and intensity of rainfall, is the wetness of the soil immediately prior to the rainfall or snow melt occurring. As described on p. 59, the amount of infiltration into a soil and subsequent storm runoff are highly dependent on the degree of saturation in the soil. Almost all major flood events are heavily influenced by the amount of

Table 5.4 Flooding events in news reports during June–July 2007

Location (date)	*Rainfall or flood statistics*	*Effect*
Midlands and Yorkshire, UK (June 2007)	1 location 103 mm of rainfall in 24 hours; many places recorded over 50 mm of rain in 12 hours	30,000+ houses affected; estimated £1.5bn damage
New South Wales, Australia (June 2007)	300 mm rainfall in 3 days	9 lives lost, 5,000 evacuated
Bangladesh (June 2007)	400 mm cumulative rainfall in places	130 lives lost, 10,000 evacuated
India (June 2007)	475 mm rain in 4 days	57 lives lost, 100,000 people evacuated
China (June–July 2007)	300 mm rainfall in 4 days	88 lives lost, 500,000 people evacuated; 56,000 homes destroyed; 91,800 ha crops destroyed
Mid-West, USA (July 2007)	305 mm rainfall in 7 days	17 lives lost
Pakistan (July 2007)	105 mm rainfall in 12 hours; 30 year record	110 lives lost, 200,000 homeless
Southern Japan (July 2007)	200 mm rainfall in 4 days	3,400 evacuated
Sudan (July 2007)	At several sites the Nile was more than a metre higher than in 1988 (a previous record level)	59 lives lost; 30,000 homes evacuated
Northland, New Zealand (July 2007)	270 mm rain total; 213 mm rain in 24 hours; 1 in 150 year storm	23 houses destroyed. Estimated damage \$80M (≈ US\$60M)
Midlands, England, UK (July 2007)	121 mm of rain in 24 hours; wettest May–July since records began in 1766	7 people killed, estimated £2bn damage

rainfall that has occurred prior to the actual flood-causing rainfall.

Deforestation

The effect of trees on runoff has already been described, particularly with respect to water resources. There is also considerable evidence that a large vegetation cover, such as forest, decreases the severity of flooding. There are several reasons for this. The first has already been described, in that trees provide an intercepting layer for rainfall and therefore slow down the rate at which the water reaches the surface. This will lessen the amount of rainfall available for soil moisture and therefore the antecedent soil moisture may be lower under forest than for an adjoining pasture (NB this is not always the case, it is dependent on the time of year). The second factor is that forests often have a high organic matter in the upper soil layers which, as any gardener will tell you, is able to absorb more water. Again this lessens the amount of overland flow, although it may increase the amount of throughflow. Finally, the infiltration rates under forest soils are often higher, leading again to less saturation excess overland flow.

The removal of forests from a catchment area will increase the propensity for a river to flood and also increase the severity of a flood event. Conversely the planting of forests on a catchment area will decrease the frequency and magnitude of flood events. Fahey and Jackson (1997) show that after conversion of native tussock grassland to exotic pine plantations a catchment in New Zealand showed a decrease in the mean flood peaks of 55–65 per cent. Although data of this type look alarming they are almost always taken from measurements at the small research catchment scale. At the larger scale the influence of deforestation is much harder to detect (see Chapter 8).

Urbanisation

Urban areas have a greater extent of impervious surfaces than in most natural landforms. Consequently the amount of infiltration excess (Hortonian) overland flow is high. In addition to this, urban areas are often designed to have a rapid drainage system, taking the overland flow away from its source. This network of gutters and drains frequently leads directly to a river drainage system, delivering more flood water in a faster time. Where extensive urbanisation of a catchment occurs; flood frequency and magnitude increases. Cherkauer (1975) shows a massive increase in flood magnitude for an urban catchment in Wisconsin, USA when compared to a similar rural catchment (see pp. 169–170). Urbanisation is another influence on flooding that is most noticeable at the small scale. This is mostly because the actual percentage area covered by impermeable urban areas in a larger river catchment is still very small in relation to the amount of permeable non-modified surfaces.

River channel alterations

Geomorphologists traditionally view a natural river channel as being in equilibrium with the river flowing within it. This does not mean that a natural river channel never floods, but rather that the channel has adjusted in shape in response to the normal discharge expected to flow through it. When the river channel is altered in some way it can have a detrimental effect on the flood characteristics for the river. In particular, **channelisation** using rigid structures can increase flood risk. Ironically, channelisation is often carried out to lessen flood risk in a particular area. This is frequently achieved, but in doing so water is passed on downstream at a faster rate than normal, increasing the flood risk further downstream. If there is a natural floodplain further downstream this may not be a problem, but if there is not, downstream riparian zones will be at greater risk.

Land drainage

It is common practice in many regions of the world to increase agricultural production through the drainage of ‘swamp’ areas. During the seventeenth and eighteenth centuries huge areas of the fenlands of East Anglia in England were drained and now are highly productive cereal and horticultural areas. The drainage of these regions provides for rapid removal of any surplus water, i.e. not needed by plants. Drained land will be drier than might be expected naturally, and therefore less storm runoff might be assumed. This is true in small rainfall events but the rapid removal of water through subsurface and surface drainage leads to flood peaks in the river drainage system where normally the water would have been slower to leave the land surface. So, although the drainage of land leads to an overall drying out of the affected area it can also lead to increased flooding through rapid drainage. Again this is scale-related, as described further in Chapter 8.

Climate change

In recent years any flooding event has led to a clamour of calls to explain the event in terms of climatic change. This is not easy to do as climate is naturally so variable. What can be said though is that river channels slowly adjust to changes in flow regime which may in turn be influenced by changes

Case study

MOZAMBIQUE FLOODS OF 2000

During the early months of 2000 world news was dominated by the catastrophic flooding that occurred in southern Africa and Mozambique in particular. The most poignant image from this time was the rescuing of a young mother, Sophia Pedro, with her baby Rosita, born up a tree while they sought refuge from the flood waters. The international media coverage of the devastating flood damage and the rescue operation that followed has ensured that this flood will be remembered for a long time to come. It has given people the world over a reminder that flooding is a hydrological hazard capable of spreading devastation on a huge scale.

The floods of Mozambique were caused by four storms in succession from January through to March 2000. The first three months of the year are the rainy season (or monsoon) for south-eastern Africa and it is usual for flooding to occur, although not to the scale witnessed in 2000. The monsoon started early in southern Mozambique; the rainfall in Maputo was 70 per cent above normal for October–November 1999. This meant that any heavy rainfall later in the rainy season would be more likely to cause a flood.

The first flood occurred during January 2000 when the Incomáti and Maputo rivers (see Figure 5.14) both burst their banks causing widespread disruption. The second flood occurred in early February, as the waters started to recede, except that now Cyclone Connie brought record rainfall to southern Mozambique and northern South Africa. The Limpopo river was as high as ever recorded (previous high was in 1977) and major communication lines were cut. The third flood, 21 February until the end of February, occurred when Cyclone Eline moved inland giving record rainfall in Zimbabwe and northern South Africa, causing record-breaking floods. The Limpopo was 3 m higher than any recorded flood and for the first time in recorded history the Limpopo and Incomáti rivers joined together in a huge inundation. The extent of the flooding can be seen in the satellite images (see Plates 7 and 8). The fourth flood was similar in size to the second and occurred following Cyclone Glória in early March (Christie and Hanlon, 2001).

There is no doubt that the Mozambique floods were large and catastrophic. How large they are, in terms of return periods or average recurrence intervals (see Chapter 6) is difficult to assess. The major difficulty is to do with paucity of streamflow records and problems with measuring flows during flood events. On the Incomáti river the flow records go back to 1937, and this was the largest flood recorded. For the Limpopo there is some data back to the 1890s, and again this was the largest recorded flow event. On the Maputo river to the south the flood levels were slightly lower than a 1984 event. The difficulties in measuring riverflow during large flood events is well illustrated by the failure of many gauging stations to function properly, either through complete inundation or being washed away. Christie and Hanlon (2001) quote an estimate of the flood on the Limpopo having a 100-year average recurrence interval, although this is difficult to verify as most gauges failed. Smithers *et al*. (2001) quote an unpublished report by Van Bladeren and Van der Spuy (2000) suggesting that upstream tributaries of the Incomáti river exceeded the 100-year return period. Smithers *et al*. (2001) provide an analysis of the 1–7 day rainfall for the Sabie catchment (a tributary of the Incomáti) which shows that in places the 200-year return period was exceeded. (NB this is an analysis of rainfall records not riverflow.)

The reasons for the flooding were simple, as they are in most cases: there was too much rainfall for the river systems to cope with the resultant

Figure 5.14 Location of the Incomáti, Limpopo and Maputo rivers in southern Africa.

stormflow. The river catchments were extremely wet (i.e. high antecedent soil moisture values) prior to the extreme rainfall, due to a prolonged and wet monsoon. One possible explanation for the severity of the rainfall is linked in with the ENSO (El Niño: Southern Oscillation) ocean-weather patterns in the Pacific. Christie and Hanlon (2001) present evidence that during a La Niña event (extreme cold temperatures in the western Pacific Ocean) it is common to see higher rainfall totals in Mozambique. However, this is not a strong relationship and certainly could not be used to make predictions. Figure 5.15 shows the monsoon rainfall at Maputo (averaged over two rainy seasons) and associated La Niña events. There may be some link here but it is not immediately obvious, particularly when you consider 1965–66 which had high rainfall despite it being an El Niño event (often associated with drought in southern Africa).

What was unusual about the 2000 floods was that the tropical Eline cyclone (called typhoons or hurricanes elsewhere) moved inland, taking extremely high rainfall to Zimbabwe and northern

South Africa. This is not normal behaviour for this type of storm and in so doing it created large floods in the headwaters of rivers draining into Mozambique. Flood warnings were issued by Zimbabwe and South Africa but the poor state of communications in Mozambique (exacerbated by the previous floods cutting communication lines) meant that they were not available to warn people on the ground. In all 700 people died as a result of the floods and 45,000 people were displaced. It is estimated that it will cost US$450 million to repair damage to the infrastructure in Mozambique (Christie and Hanlon, 2001). This is not the total cost of the flood, which is far higher when loss of income and loss of private property are included. These costs will never be fully known as in many lesser-developed countries the costs are borne by individuals without any form of insurance cover.

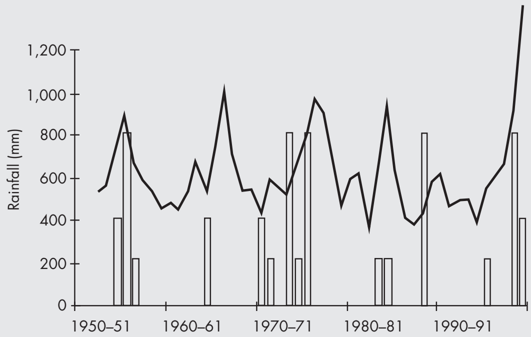

Figure 5.15 Rainfall totals during the rainy season (smoothed with a two-year average) at Maputo airport, with vertical bars indicating the strength of La Niña events (on a scale of three: strong, medium, weak).

Sources: Rainfall data from Christie and Hanlon (2001); La Niña strength from NOAA

In many ways there are no new lessons to learn from the Mozambique floods of 2000. It is well known that adequate warning systems are needed (but expensive) and that people should be restricted from living in flood-prone areas; but this is difficult to achieve in a poor country such as Mozambique. The cause of the flood was a huge amount of rainfall and the severity was influenced by the antecedent wetness of the ground due to a very wet monsoon.

in climate. Many studies have suggested that future climate change will involve greater extremes of weather (IPCC, 2007), including more high-intensity rainfall events. This is likely to lead to an increase in flooding, particularly while a channel adjusts to the differing flow regime (if it is allowed to).

RUNOFF IN THE CONTEXT OF WATER QUALITY

The route that water takes between falling as precipitation and reaching a stream has a large influence on water quality. The nutrient level of water is heavily influenced by the length of time water spends in contact with soil. Water that moves quickly into a river (e.g. overland flow) is likely to have a lower nutrient level than water that moves slowly through the soil as throughflow and/or groundwater. However, water that has travelled as overland flow may have a higher level of suspended solids picked up from the surface, so it may appear less pure.

In considering issues of land-use change and water quality, an important consideration is the time taken for water to reach the stream. It is important to realise that groundwater is frequently operating as a pressure wave response to rainfall recharge. Where groundwater responds to a rainfall event by emitting water into a stream it is a pressure wave response, i.e. the water entering the stream is not the same water that infiltrates and causes the response. This means that water entering the stream may be several years (or more) older and unaffected by the current land use change.

SUMMARY

The water flowing down a river is the end-product of precipitation after all the other hydrological processes have been in operation. The sub-processes of overland flow, throughflow and groundwater flow are well understood, although it is not easy to estimate their relative importance for a particular site, particularly during a storm event. The measurement of river flow is relatively straightforward and presents the fewest difficulties in terms of sampling error, although there are limitations, particularly during periods of high flow and floods.

FURTHER READING

Anderson, M.G. and Burt, T.P. (eds) (1990) *Process studies in hillslope hydrology*. Wiley, Chichester.
A slightly more modern update on Kirkby (1978).

Kirkby, M.J. (ed.) (1978) *Hillslope hydrology*. J. Wiley & Sons, Chichester.
A classic text on hillslope processes, particularly runoff.

Parsons, A.J. and Abrahams, A.D. (1992) *Overland flow: hydraulics and erosion mechanics*. UCL Press, London.
An advanced edited book; good detail on arid regions.

Smith, K. and Ward, R.C. (1998) *Floods: physical processes and human impacts*. Wiley, Chichester.
A text on flooding.

Wohl, E.E. (2000) *Inland flood hazards: human, aquatic and riparian communities*. Cambridge University Press, Cambridge.
A text on flooding with many case studies.

6

STREAMFLOW ANALYSIS AND MODELLING

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of what different hydrological techniques are used for.
- A knowledge of hydrograph analysis (including the unit hydrograph).
- A knowledge of how to derive and interpret flow duration curves.
- A knowledge of how to carry out frequency analysis, particularly for floods.
- An understanding of the aims of hydrological modelling and different strategies to achieve those aims.

One of the most important tasks in hydrology is to analyse streamflow data. These data are continuous records of discharge, frequently measured in permanent structures such as flumes and weirs (see Chapter 5). Analysis of these data provides us with three important features:

- description of a flow regime
- potential for comparison between rivers, and
- prediction of possible future river flows.

There are well-established techniques available to achieve these, although they are not universally applied in the same manner. This chapter sets out three important methods of analysing streamflow: hydrograph analysis, flow duration curves and frequency analysis. These three techniques are explained with reference to worked examples, all drawn from the same data set. The use of data from within the same study area is important for comparison between the techniques.

HYDROGRAPH ANALYSIS

A hydrograph is a continuous record of stream or river discharge (see Figure 5.1). It is a basic working unit for a hydrologist to understand and interpret. The shape of a hydrograph is a response from a particular catchment to a series of unique conditions, ranging from the underlying geology and catchment shape to the antecedent wetness and storm duration. The temporal and spatial variations in these underlying conditions make it highly unlikely that two hydrographs will ever be the same. Although there is great variation in the shape of a hydrograph there are common characteristics of a storm hydrograph that can be recognised. These have been described at the start of Chapter 5 where terms such as *rising limb*, *falling limb*, *recession limb* and *baseflow* are explained.

Hydrograph separation

The separation of a hydrograph into baseflow and stormflow is a common task, although never easy. The idea of **hydrograph separation** is to distinguish between stormflow and baseflow so that the amount of water resulting from a storm can be calculated. Sometimes further assumptions are made concerning where the water in each component has come from (i.e. groundwater and overland flow) but, as explained in the previous chapter, this is controversial.

The simplest form of hydrograph separation is to draw a straight, level line from the point where the hydrograph starts rising until the stream discharge reaches the same level again (dashed line in Figure 6.1). However, this is frequently problematic as the stream may not return to its pre-storm level before another storm arrives. Equally the storm may recharge the baseflow enough so that the level is raised after the storm (as shown in Figure 6.1).

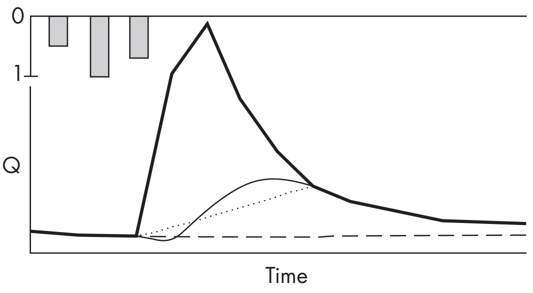

Figure 6.1 Hydrograph separation techniques. See text for explanation.

To overcome the problem of a level baseflow separation a point has to be chosen on the receding limb where it is decided that the discharge has returned to baseflow. Exactly where this point will be is not easy to determine. By convention the point is taken where the recession limb fits an exponential curve. This can be detected by plotting the natural log (ln) of discharge (Q) and noting where this line becomes straight. The line drawn between the start and 'end' of a storm may be straight (dotted line, see Figure 6.1) or curved (thin solid line, see Figure 6.1) depending on the preference of hydrologist – Arnold *et al*. (1995) provides a summary of different automated techniques.

In very large catchments equation 6.1 can be applied to derive the time where stormflow ends. This is the fixed time method which gives the time from peak flow to the end of stormflow (τ):

$$\tau = D^n \tag{6.1}$$

where D is the drainage area and n is a recession constant. When D is in square miles and τ in days, the value of n has been found to be approximately 0.2.

The problem with hydrograph separation is that the technique is highly subjective. There is no physical reasoning why the 'end' of a storm should be when the recession limb fits an exponential curve; it is pure convention. Equally the shape of the curve between start and 'end' has no physical reasoning. It does not address the debate covered in Chapter 5: where does the stormflow water come from? Furey and Gupta (2001) have recently provided a hydrograph separation technique that ties into physical characteristics of a catchment and therefore is not as subjective as other techniques, although it still requires considerable interpretation by the user. What hydrograph separation does offer is a means of separating stormflow from baseflow, something that is needed for the use of the unit hydrograph (see pp. 103–106), and may be useful for hydrological interpretation and description.

The unit hydrograph

The idea of a **unit hydrograph** was first put forward by Sherman, an American engineer working in the 1920s and 1930s. The idea behind the unit hydrograph is simple enough, although it is a somewhat tedious exercise to derive one for a catchment. The fundamental concept of the unit hydrograph is that the shape of a storm hydrograph is determined by the physical characteristics of the catchment. The majority of those physical characteristics are static in time, therefore if you can find an average hydrograph for a particular storm size then you can use that to predict other storm events. In short: two identical rainfall events that fall on a catchment with exactly the same antecedent conditions should produce identical hydrographs.

With the unit hydrograph a hydrologist is trying to predict a future storm hydrograph that will result from a particular storm. This is particularly useful as it gives more than just the peak runoff volume and includes the temporal variation in discharge.

Sherman (1932) defines a unit hydrograph as 'the hydrograph of surface runoff resulting from **effective rainfall** falling in a unit of time such as 1 hour or 1 day'. The term effective rainfall is taken to be that rainfall that contributes to the storm hydrograph. This is often assumed to be the rainfall that does not infiltrate the soil and moves into the stream as overland flow. This is infiltration excess or Hortonian overland flow. Sherman's ideas fitted well with those of Horton. Sherman assumed that the 'surface runoff is produced uniformly in space and time over the total catchment area'.

Deriving the unit hydrograph: step 1

Take historical rainfall and streamflow records for a catchment and separate out a selection of typical single-peaked storm hydrographs. It is important that they are separate storms as the method assumes that one runoff event does not affect another. For each of these storm events separate the baseflow from the stormflow; that is, hydrograph separation (see p. 102). This will give you a series of storm hydrographs (without the baseflow component) for a corresponding rainfall event.

Deriving the unit hydrograph: step 2

Take a single storm hydrograph and find out the total volume of water that contributed to the storm. This can be done either by measuring the area under the stormflow hydrograph or as an integral of the curve. If you then divide the total volume in the storm by the catchment area, you have the runoff as a water equivalent depth. If this is assumed to have occurred uniformly over space and time within the catchment then you can assume it is equal to the effective rainfall. This is an important assumption of the method: that the effective rainfall is equal to the water equivalent depth of storm runoff. It is also assumed that the effective rainfall all occurred during the height of the storm (i.e. during the period of highest rainfall intensity). That period of high rainfall intensity becomes the time for the unit hydrograph.

Deriving the unit hydrograph: step 3

The unit hydrograph is the stormflow that results from one unit of effective rainfall. To derive this you need to divide the values of stormflow (i.e. each value on the storm hydrograph) by the amount of effective rainfall (from step 2) to give the unit hydrograph. This is the discharge per millimetre of effective rainfall during the time unit.

Deriving the unit hydrograph: step 4

Repeat steps 2 and 3 for all of the typical hydrographs. Then create an average unit hydrograph by merging the curves together. This is achieved by averaging the value of stormflow for each unit of time for every derived unit hydrograph. It is also possible to derive different unit hydrographs for different rain durations and intensities, but this is not covered here (see Maidment, 1992, or Shaw, 1994, for more details).

Using the unit hydrograph

The unit hydrograph obtained from the steps described here theoretically gives you the runoff that can be expected per mm of effective rainfall in one hour. In order to use the unit hydrograph for predicting a storm it is necessary to estimate the 'effective rainfall' that will result from the storm rainfall. This is not an easy task and is one of the main hurdles in using the method. In deriving the unit hydrograph the assumption has been made that 'effective rainfall' is the rainfall which does not infiltrate but is routed to the stream as overland flow (Hortonian). The same assumption has to be made when utilising the unit hydrograph. To do this it is necessary to have some indication of the infiltration characteristics for the catchment concerned, and also of the antecedent soil moisture conditions. The former can be achieved through field experimentation and the latter through the use of an antecedent precipitation index (API). Engineering textbooks give examples of how to use the API to derive effective rainfall. The idea is that antecedent soil moisture is controlled by how long ago rain has fallen and how large that event was. The wetter a catchment is prior to a storm, the more effective rainfall will be produced.

Once the effective rainfall has been established it is a relatively simple task to add the resultant unit hydrographs together to form the resultant storm hydrograph. The worked example shows how this procedure is carried out.

Limitations of the unit hydrograph

The unit hydrograph has several assumptions that at first appearance would seem to make it inapplicable in many situations. The assumptions can be summarised as:

- The runoff that makes up stormflow is derived from infiltration excess (Hortonian) overland flow. As described in Chapter 5, this is not a reasonable assumption to make in many areas of the world.
- That the surface runoff occurs uniformly over the catchment because the rainfall is uniform over the catchment. Another assumption that is difficult to justify.
- The relationship between effective rainfall and surface runoff does not vary with time (i.e. the hydrograph shape remains the same between the data period of derivation and prediction). This would assume no land-use change within the catchment, as this could well affect the storm hydrograph shape.

Given the assumptions listed above it would seem extremely foolhardy to use the unit hydrograph as a predictive tool. However, the unit hydrograph has been used successfully for many years in numerous different hydrological situations. It is a very simple method of deriving a storm hydrograph from a relatively small amount of data. The fact that it does work (i.e. produces meaningful predictions of storm hydrographs), despite being theoretically flawed, would seem to raise questions about our understanding of hydrological processes. The answer to why it works may well lie in the way that it is applied, especially the use of effective rainfall. This is a nebulous concept that is difficult to describe from field measurements. It is possible that in moving from actual to effective rainfall there is a blurring of processes that discounts some of the assumptions listed above. The unit hydrograph is a black box model of stormflow (see end of this chapter) and as such hides many different processes within. The simple concept that the hydrograph shape is a reflection of the static characteristics and all the dynamic processes going on in a catchment makes it highly applicable but less able to be explained in terms of hydrological theory.

The synthetic unit hydrograph

The **synthetic unit hydrograph** is an attempt to derive the unit hydrograph from measurable catchment characteristics rather than from flow data. This is highly desirable as it would give the opportunity to predict stormflows when having no

Worked example of the unit hydrograph

The Tanllwyth is a small (0.98 km^2) headwater tributary of the Severn river in mid-Wales. The catchment is monitored by the Centre for Ecology and Hydrology (formerly the Institute of Hydrology) as part of the Plynlimon catchment experiment. For this example a storm was chosen from July 1982 as it is a simple single-peaked hydrograph (see Figure 6.2).

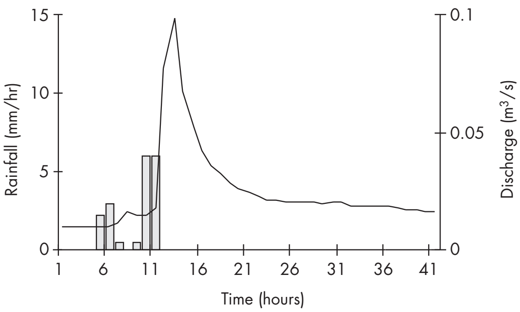

Figure 6.2 A simple storm hydrograph (July 1982) from the Tanllwyth catchment.

Baseflow separation was carried out by using a straight-line method. The right-hand end of the straight line (shown as a broken line in Figure 6.3) is where the receding limb of the hydrograph became exponential.

All of the flow above the broken line in Figure 6.3 was then divided by the effective rainfall to derive the unit hydrograph in Figure 6.4 below.

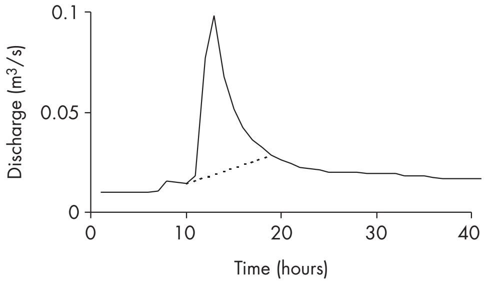

Figure 6.3 Baseflow separation.

Figure 6.4 The unit hydrograph for the Tanllwyth catchment.

NB the hydrograph appears more spread out because of the scale of drawing.

To apply the unit hydrograph to a small storm, hydrographs were added together for each amount of effective rainfall. The resultant total hydrograph is shown as a dark black line in Figure 6.5. The discharge values in the simulated hydrograph are much larger than those in the original storm hydrograph despite what appears to be a smaller storm. This is because the simulated hydrograph is working on effective rainfall rather than actual rainfall. Effective rainfall is the rain that doesn't infiltrate and is theoretically available for storm runoff. A low effective rainfall value may represent a high actual rainfall value.

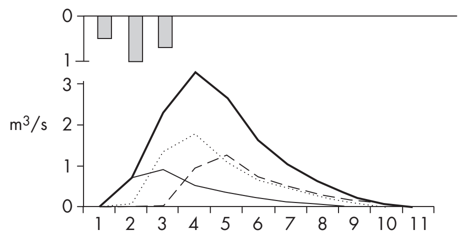

Figure 6.5 Applying the unit hydrograph to a small storm (effective rainfall shown on the separate scale above). The different lines represent the flow from each of the rainfall bars (thin solid first, then dotted, then dashed). The solid black line is the total discharge i.e. the sum of the three lines.

historical streamflow data; a common predicament around the world. The Institute of Hydrology in the UK carried out an extensive study into producing synthetic unit hydrographs for catchments, based on factors such as the catchment size, degree of urbanisation, etc. (NERC, 1975). They produced a series of multiple regression equations to predict peak runoff amount, time to peak flow, and the time to the end of the recession limb based on the measurable characteristics. Although this has been carried out relatively successfully it is only applicable to the UK as that is where the derivative data was from. In another climatic area the hydrological response is likely to be different for a similar catchment. The UK is a relatively homogeneous climatic area with a dense network of river flow gauging, which allowed the study to be carried out. In areas of the world with great heterogeneity in climate and sparse river monitoring it would be extremely difficult to use this approach.

FLOW DURATION CURVES

An understanding of how much water is flowing down a river is fundamental to hydrology. Of particular interest for both flood and low flow hydrology is the question of how representative a certain flow is. This can be addressed by looking at the frequency of daily flows and some statistics that can be derived from the frequency analysis. The culmination of the frequency analysis is a **flow duration curve** which is described below.

Flow duration curves are concerned with the amount of time a certain flow is exceeded. The data most commonly used are daily mean flows: the average flow for each day (note well that this is not the same as a mean daily flow, which is the average of a series of daily flows). To derive a flow duration curve the daily mean flow data are required for a long period of time, in excess of five years. A worked example is provided here, using twenty-six years of data for the upper reaches of the river Wye in mid-Wales, UK (see pp. 108–109).

Flow duration curve: step 1

A table is derived that has the frequency, cumulative frequency (frequency divided by the total number of observations) and percentage cumulative frequency. The percentage cumulative frequency is assumed to equal the percentage of time that the flow is exceeded. While carrying out the frequency analysis it is important that a small class (or bin) interval is used; too large an interval and information will be lost from the flow duration curve. The method for choosing the best class interval is essentially through trial and error. As a general rule you should aim not to have more than around 10 per cent of your values within a single class interval. If you have more than this you start to lose precision in plotting. As shown in the worked example, it is not essential that the same interval is used throughout.

Flow duration curve: step 2

The actual flow duration curve is created by plotting the percentage cumulative frequency on the x-axis against the mid-point of the class interval on the y-axis. Where two flow duration curves are presented on the same axes they need to be standardised for direct comparison. To do this the values on the y-axis (mid-point of class interval) are divided by the average flow for the record length. This makes the y-axis a percentage of the average flow (see Figure 6.6).

The presentation of a flow duration curve may be improved by either plotting on a special type of graph paper or transforming the data. The type of graph paper often used has the x-axis transformed in the form of a known distribution such as the Gumbel or Log Pearson. A natural log transformation of the flow values (y-axis) achieves a similar effect, although this is not necessarily standard practice.

Interpreting a flow duration curve

The shape of a flow duration curve can tell a lot about the hydrological regime of a catchment. In Figure 6.6 two flow duration curves of contrasting shape are shown. With the dotted line there is a

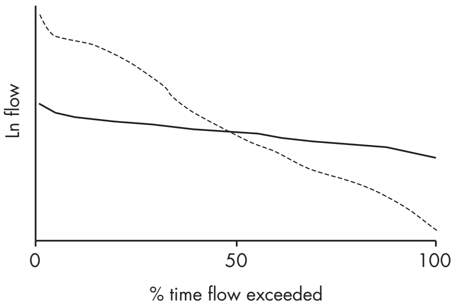

Figure 6.6 Two contrasting flow duration curves. The dotted line has a high variability in flow (similar to a small upland catchment) compared to the solid line (similar to a catchment with a high baseflow).

large difference between the highest and lowest flow values, whereas for the solid line there is far less variation. This tells us that the catchment shown by the solid line never has particularly low flows or particularly high flows. This type of hydrological response is found in limestone or chalk catchments where there is a high baseflow in the summer (groundwater derived) and high infiltration rates during storm events. In contrast the catchment shown with a solid line has far more variation. During dry periods it has a very low flow, but responds to rainfall events with a high flow. This is characteristic of impermeable upland catchments or streamflow in dryland areas.

Statistics derived from a flow duration curve

The interpretation of flow duration curve shape discussed so far is essentially subjective. In order to introduce some objectivity there are statistics derived from the curve; the three most important ones are:

- The flow value that is exceeded 95 per cent of the time (Q_{95}). A useful statistic for low flow analysis.
- The flow value that is exceeded 50 per cent of the time (Q_{50}). This is the median flow value.
- The flow value that is exceeded 10 per cent of the time (Q_{10}). A useful statistic for analysis of high flows and flooding.

FREQUENCY ANALYSIS

The analysis of how often an event is likely to occur is an important concept in hydrology. It is the application of statistical theory into an area that affects many people's lives, whether it be through flooding or low flows and drought. Both of these are considered here, although because they use similar techniques the main emphasis is on **flood frequency analysis**. The technique is a statistical examination of the frequency–magnitude relationship discussed in Chapters 2 and 5. It is an attempt to place a probability on the likelihood of a certain event occurring. Predominantly it is concerned with the low-frequency, high-magnitude events (e.g. a large flood or a very low river flow).

It is important to differentiate between the uses of flow duration curves and frequency analysis. Flow duration curves tell us the percentage of time that a flow is above or below a certain level. This is average data and describes the overall flow regime. Flood frequency analysis is concerned only with peak flows: the probability of a certain flood recurring. Conversely, **low flow frequency analysis** is concerned purely with the lowest flows and the probability of them recurring.

Flood frequency analysis

Flood frequency analysis is probably the most important hydrological technique. The concept of a '100-year flood', or a fifty-year recurrence interval, is well established in most people's perceptions of hydrology, although there are many misunderstandings in interpretation.

Flood frequency analysis is concerned with peak flows. There are two different ways that a peak flow can be defined:

- the single maximum peak within a year of record giving an **annual maximum series**; or
- any flow above a certain threshold value, giving a **partial duration series**.

Figure 6.10 shows the difference between these two data series. There are arguments for and against the

Worked example of flow duration curve

The Wye river has its headwaters in central Wales and flows into the Severn at the head of the Severn Estuary. In its upper reaches it is part of the Plynlimon hydrological experiment run by the Institute of Hydrology (now the Centre for Ecology and Hydrology) from the early 1970s. At Plynlimon the Wye is a small (10.5 km^2) grassland catchment with an underlying geology of relatively impermeable Ordovician shale. The data used to derive a flow duration curve here are from the Upper Wye for a period from 1970 until 1995, consisting of 9,437 values of daily mean flow in cumecs.

The frequency analysis for the Wye gives Table 6.1 opposite.

The flow duration curve is derived by plotting the percentage cumulative frequency (x-axis) against the mid-point of the daily mean flow class intervals (y-axis). When this is plotted it forms the exponential shape that is normal for this type of catchment (see Figure 6.7). In order to see more detail on the curve the flow values can be logged (natural log). This is shown in Figure 6.8

The flow statistics Q_{95}, Q_{50} and Q_{10} can either be read from the graph (see Figure 6.9) or interpolated from the original frequency table (remembering to use the mid-points of the class interval).

A summary of the flow statistics for the upper Wye are shown in Table 6.2 below.

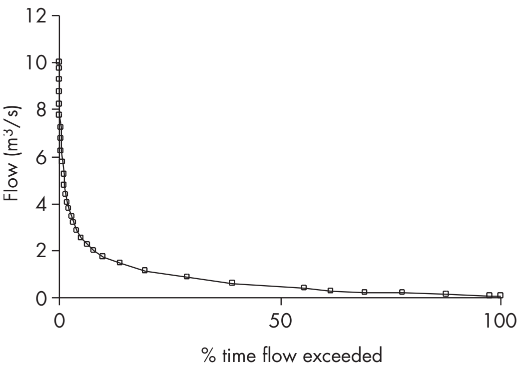

Figure 6.7 Flow duration curve for the river Wye (1970–1995 data).

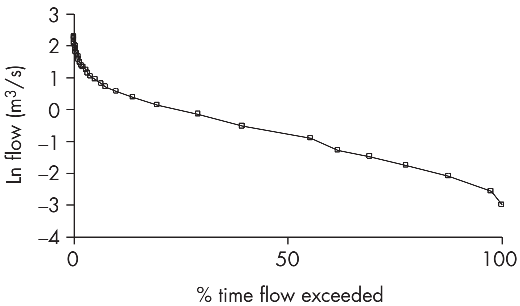

Figure 6.8 Flow duration curve for the river Wye (1970–1995 data) with the flow data shown on a natural log scale.

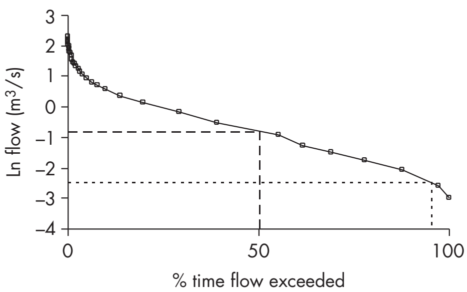

Figure 6.9 Q_{95} (short dashes) and Q_{50} (long dashes) shown on the flow duration curve.

Table 6.1 Values from the frequency analysis of daily mean flow on the upper Wye catchment. These values form the basis of the flow duration curve in Figure 6.7

Daily mean flow (m^3/s)	*Frequency*	*Relative frequency (%)*	*Cumulative frequency (%)*
0–0.05	250	2.65	100.00
0.05–0.1	923	9.78	97.35
0.1–0.15	927	9.82	87.57
0.15–0.2	814	8.63	77.75
0.2–0.25	708	7.50	69.12
0.25–0.3	589	6.24	61.62
0.3–0.4	881	9.34	55.38
0.4–0.5	641	6.79	46.04
0.5–0.7	958	10.15	39.25
0.7–1.0	896	9.49	29.10
1.0–1.3	553	5.86	19.60
1.3–1.6	357	3.78	13.74
1.6–1.9	222	2.35	9.96
1.9–2.1	117	1.24	7.61
2.1–2.4	127	1.35	6.37
2.4–2.7	103	1.09	5.02
2.7–3.0	71	0.75	3.93
3.0–3.3	47	0.50	3.18
3.3–3.6	42	0.45	2.68
3.6–3.9	34	0.36	2.24
3.9–4.2	33	0.35	1.88
4.2–4.5	28	0.30	1.53
4.5–5.0	28	0.30	1.23
5.0–5.5	23	0.24	0.93
5.5–6.0	23	0.24	0.69
6.0–6.5	14	0.15	0.45
6.5–7.0	7	0.07	0.30
7.0–7.5	7	0.07	0.22
7.5–8.0	3	0.03	0.15
8.0–8.5	4	0.04	0.12
8.5–9.0	2	0.02	0.07
9.0–9.5	1	0.01	0.05
9.5–10.0	3	0.03	0.04
> 10.0	1	0.01	0.01
Total	9,437	100	

Table 6.2 Summary flow statistics derived from the flow duration curve for the Wye catchment

Flow statistic	*Ln flow (m^3/s)*	*Flow (m^3/s)*
Q_{95}	–2.48	0.084
Q_{50}	–0.78	0.458
Q_{10}	0.56	1.75

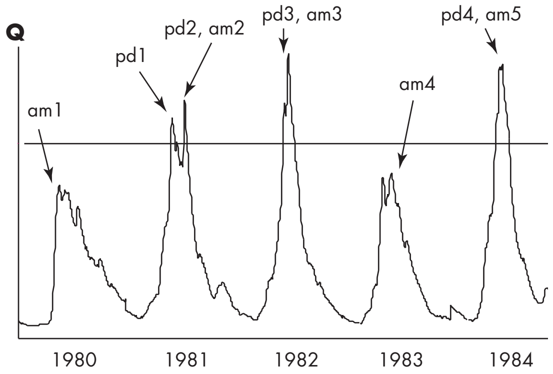

Figure 6.10 Daily flow record for the Adams river (British Columbia, Canada) during five years in the 1980s. Annual maximum series are denoted by 'am', partial duration series above the threshold line by 'pd'. NB In this record there are five annual maximum data points and only four partial duration points, including two from within 1981.

Source: Data courtesy of Environment Canada

use of either data series in flood frequency analysis. Annual maximum may miss a large storm event where it occurs more than once during a year (as in the 1981 case in Figure 6.10), but it does provide a continuous series of data that are relatively easy to process. The setting of a threshold storm (the horizontal line in Figure 6.10) is critical in analysis of the partial duration series, something that requires considerable experience to get right. The most common analysis is on annual maximum series, the simplest form, which is described here. If the data series is longer than ten years then the annual maxima can be used; for very short periods of record the partial duration series can be used.

The first step in carrying out flood frequency analysis is to obtain the data series (in this case annual maxima). The annual maximum series should be for as long as the data record allows. The greater the length of record the more certainty can be attached to the prediction of average recurrence interval. Many hydrological database software packages (e.g. HYDSYS) will give annual maxima data automatically, but some forethought is required on what annual period is to be used. There is an assumption made in flood frequency analysis that the peak flows are independent of each other (i.e. they are not part of the same storm). If a calendar year is chosen for a humid temperate environment in the northern hemisphere, or a tropical region, it is possible that the maximum river flow will occur in the transition between years (i.e. December/January). It is possible for a storm to last over the 31 December/1 January period and the same storm to be the maximum flow value for both years. If the flow regime is dominated by snow melt then it is important to avoid splitting the hydrological year at times of high melt (e.g. spring and early summer). To avoid this it is necessary to choose your hydrological year as changing during the period of lowest flow. This may take some initial investigation of the data.

All flood frequency analysis is concerned with the analysis of frequency histograms and probability distributions. Consequently the first data analysis step should be to draw a frequency histogram. It is often useful to convert the frequency into a relative frequency (divide the number of readings in each class interval by the total number of readings in the data series).

The worked example given is for a data set on the river Wye in mid-Wales (see pp. 113–114). On looking at the histogram of the Wye data set (Figure 6.11) the first obvious point to note is that it is not normally distributed (i.e. it is not a classic bell-shaped curve). It is important to grasp the significance of the non-normal distribution for two reasons:

1 Common statistical techniques that require normally distributed data (e.g. t-tests etc.) cannot be applied in flood frequency analysis.
2 It shows what you might expect: small events are more common than large floods, but that very large flood events do occur; i.e. a high magnitude, low frequency relationship.

If you were to assume that the data series is infinitely large in number and the class intervals were made extremely small, then a smooth curve can be drawn

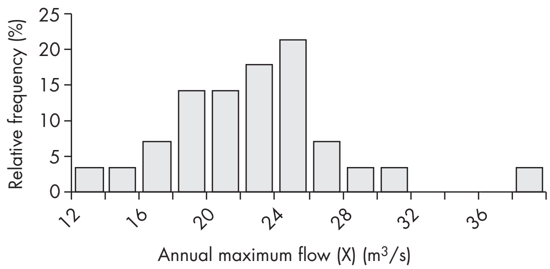

Figure 6.11 Frequency distribution of the Wye annual maximum series.

through the histogram. This is the *probability density function* which represents the smoothed version of your frequency histogram.

In flood frequency analysis there are three interrelated terms of interest. These terms are interrelated mathematically, as described in equation 6.2 in the text below.

1 The probability of excedence: $P(X)$. This is the probability that a flow (Q) is greater than, or equal to a value X. The probability is normally expressed as a unitary percentage (i.e. on a scale between 0 and 1).
2 The relative frequency: $F(X)$. This is the probability of the flow (Q) being less than a value X. This is also expressed as a unitary percentage.
3 The average recurrence interval: $T(X)$. This is sometimes referred to as the return period, although this is misleading. $T(X)$ is a statistical term meaning the chance of excedence once every T years over a long record. This should not be interpreted as meaning that is exactly how many years are likely between certain size floods.

$$P(X) = 1 - F(X)$$
$$T(X) = \frac{1}{P(X)} = \frac{1}{1 - F(X)} \tag{6.2}$$

It is possible to read the values of $F(X)$ from a cumulative probability curve; this provides the simplest method of carrying out flood frequency analyses. One difficulty with using this method is that you must choose the class intervals for the histogram carefully so that the probability density function is an accurate representation of the data. Too large an interval and the distribution may be shaped incorrectly, too small and holes in the distribution will appear.

One way of avoiding the difficulties of choosing the best class interval is to use a rank order distribution. This is often referred to as a plotting position formula.

The Weibull formula

The first step in the method is to rank your annual maximum series data from low to high. In doing this you are assuming that each data point (i.e. the maximum flood event for a particular year) is independent of any others. This means that the year that the flood occurred in becomes irrelevant.

Taking the rank value, the next step is to calculate the $F(X)$ term using equation 6.3. In this case r refers to the rank of an individual flood event (X) within the data series and N is the total number of data points (i.e. the number of years of record):

$$F(X) = \frac{r}{N+1} \tag{6.3}$$

In applying this formula there are two important points to note:

1 The value of $F(X)$ can never reach 1 (i.e. it is asymptotic towards the value 1).
2 If you rank your data from high to low (i.e. the other way around) then you will be calculating the $P(X)$ value rather than $F(X)$. This is easily rectified by using the formula linking the two.

The worked example on pp. 113–114 gives the $F(X)$, $P(X)$ and $T(X)$ for a small catchment in mid-Wales (Table 6.3).

The Weibull formula is simple to use and effective but is not always the best description of an annual maximum series data. Some users suggest that a better fit to the data is provided by the Gringorten formula (equation 6.4):

$$F(X) = \frac{r - 0.44}{N + 0.12} \quad (6.4)$$

As illustrated in the worked example, the difference between these two formulae is not great and often the use of either one is down to personal preference.

Extrapolating beyond your data set

The probabilities derived from the Weibull and Gringorten formulae give a good description of the flood frequency within the measured stream record but do not provide enough data when you need to extrapolate beyond a known time series. This is a common hydrological problem: we need to make an estimate on the size of a flood within an average recurrence interval of fifty years but only have twenty-five years of streamflow record. In order to do this you need to fit a distribution to your data. There are several different ways of doing this, the method described here uses the method of moments based on the Gumbel distribution. Other distributions that are used by hydrologists include the Log-Pearson Type III and log normal. The choice of distribution is often based on personal preference and regional bias (i.e. the distribution that seems to fit flow regimes for a particular region).

Method of moments

If you assume that the data fits a Gumbel distribution then you can use the method of moments to calculate $F(X)$ values. Moments are statistical descriptors of a data set. The first moment of a data set is the mean; the second moment the standard deviation; the third moment skewness; the fourth kurtosis. To use the formulae below (equations 6.5–6.7) you must first find the mean ($\bar{Q}$) *and standard deviation* (σ_Q) of your annual maximum data series. The symbol *e* in the equations 6.5–6.7 is the base number for natural logarithms or the exponential number (≈ 2.7183).

$$F(X) = e^{-e^{-b(X-a)}} \quad (6.5)$$

$$a = \bar{Q} - \frac{0.5772}{b} \quad (6.6)$$

$$b = \frac{\pi}{\sigma_Q \sqrt{6}} \quad (6.7)$$

With knowledge of $F(X)$ you can find $P(X)$ and the average recurrence interval ($T(X)$) for a certain size of flow: X. The formulae above can be rearranged to give you the size of flow that might be expected for a given average recurrence interval (equation 6.8):

$$X = a - \frac{1}{b}\ln\ln\left(\frac{T(X)}{T(X)-1}\right) \quad (6.8)$$

In the formula above ln represents the natural logarithm. To find the flow for a fifty-year average recurrence interval you must find the natural logarithm of (50/49) and then the natural logarithm of this result.

Using this method it is possible to find the resultant flow for a given average recurrence interval that is beyond the length of your time series. The further away from the length of your time series you move the more error is likely to be involved in the estimate. As a general rule of thumb it is considered reasonable to extrapolate up to twice the length of your streamflow record, but you should not go beyond this.

Low flow frequency analysis

Where frequency analysis is concerned with low flows rather than floods, the data required are an annual minimum series. The same problem is found as for annual maximum series: which annual year to use when you have to assume that the annual minimum flows are independent of each other. At mid-latitudes in the northern hemisphere the calendar year is the most sensible, as you would expect the lowest flows to be in the summer months (i.e. the middle of the year of record). Elsewhere an

Worked example of flood frequency analysis

The data used to illustrate the flood frequency analysis are from the same place as the flow duration curve (the upper Wye catchment in Wales, UK). In this case it is an annual maximum series for the period 1970 until 1997 (inclusive).

In order to establish the best time of year to set a cut off for the hydrological year all daily mean flows above a threshold value (4.5 m^3/s) were plotted against their day number (figure 6.12). It is clear from Figure 6.12 that high flows can occur at almost any time of the year although at the start and end of the summer (150 = 30 May; 250 = 9 September) there are slight gaps. The hydrological year from June to June is sensible to choose for this example.

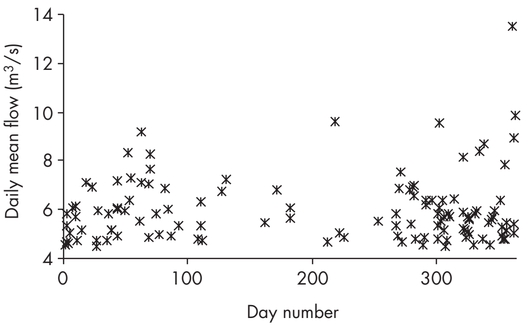

Figure 6.12 Daily mean flows above a threshold value plotted against day number (1–365) for the Wye catchment.

The Weibull and Gringorten position plotting formulae are both applied to the data (see Table 6.3) and the $F(X)$, $P(X)$ and $T(X)$ (average recurrence interval) values calculated. The data look different from those in Figure 6.12 and from the flow duration curve because they are the peak flow values recorded in each year. This is the peak value of each storm hydrograph, which is not the same as the peak mean daily flow values.

When the Weibull and Gringorten values are plotted together (Figure 6.13) it can be seen that there is very little difference between them.

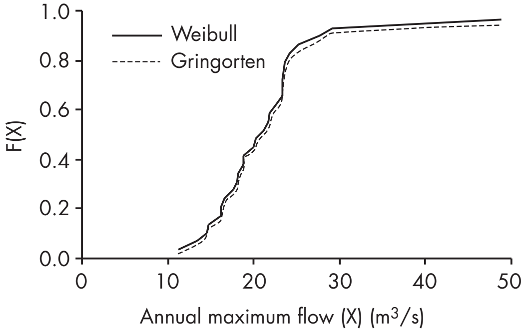

Figure 6.13 Frequency of flows less than X plotted against the X values. The *F*(X) values are calculated using both the Weibull and Gringorten formulae.

When the data are plotted with a transformation to fit the Gumbel distribution they almost fit a straight line, suggesting that they do fit a distribution for extreme values such as the Gumbel but that a larger data set would be required to make an absolute straight line. A longer period of records is likely to make the extreme outlier lie further along the x-axis. The plot presented here has transformed the data to fit the Gumbel distribution. Another method of presenting this data is to plot them on Gumbel distribution paper. This provides a non-linear scale for the x-axis based on the Gumbel distribution.

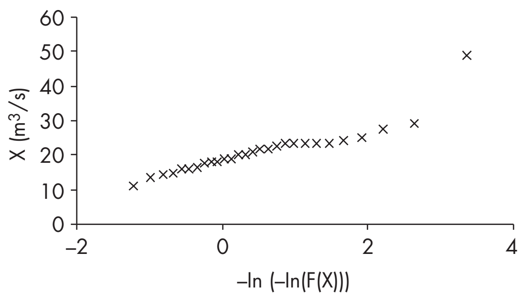

Figure 6.14 Frequency of flows less than a value X. NB The *F*(X) values on the x-axis have undergone a transformation to fit the Gumbel distribution (see text for explanation).

Table 6.3 Annual maximum series for the Wye (1971–97) sorted using the Weibull and Gringorten position plotting formulae

Rank	*X*	*F(X) Weibull*	*F(X) Gringorten*	*P(X)*	*T(X)*
1	11.17	0.03	0.02	0.97	1.04
2	13.45	0.07	0.05	0.93	1.07
3	14.53	0.10	0.09	0.90	1.12
4	14.72	0.14	0.12	0.86	1.16
5	16.19	0.17	0.16	0.83	1.21
6	16.19	0.21	0.19	0.79	1.26
7	16.58	0.24	0.23	0.76	1.32
8	17.57	0.28	0.26	0.72	1.38
9	18.09	0.31	0.29	0.69	1.45
10	18.25	0.34	0.33	0.66	1.53
11	18.75	0.38	0.36	0.62	1.61
12	18.79	0.41	0.40	0.59	1.71
13	20.01	0.45	0.43	0.55	1.81
14	20.22	0.48	0.47	0.52	1.93
15	21.10	0.52	0.50	0.48	2.07
16	21.75	0.55	0.53	0.45	2.23
17	21.84	0.59	0.57	0.41	2.42
18	22.64	0.62	0.60	0.38	2.64
19	23.28	0.66	0.64	0.34	2.90
20	23.36	0.69	0.67	0.31	3.22
21	23.37	0.72	0.71	0.28	3.63
22	23.46	0.76	0.74	0.24	4.14
23	23.60	0.79	0.77	0.21	4.83
24	24.23	0.83	0.81	0.17	5.80
25	25.19	0.86	0.84	0.14	7.25
26	27.68	0.90	0.88	0.10	9.67
27	29.15	0.93	0.91	0.07	14.50
28	48.87	0.97	0.95	0.03	29.00

Table 6.4 Values required for the Gumbel formula, derived from the Wye data set in Table 6.3

Mean ($\bar{Q}$)	*Standard deviation (σ_Q)*	*a value*	*b value*
21.21	6.91	18.11	0.19

Applying the method of moments and Gumbel formula to the data give some interesting results. The values used in the formula are shown below and can be easily computed. When the formula is applied to find the flow values for an average recurrence interval of fifty years it is calculated as 39.1 m^3/s. This is less than the largest flow during the record which under the Weibull formula has an average recurrence interval of twenty-seven years. This discrepancy is due to the method of moments formula treating the highest flow as an extreme outlier. If we invert the formula we can calculate that a flood with a flow of 48.87 m^3/s (the largest on record) has an average recurrence interval of around three hundred years.

analysis of when low flows occur needs to be carried out so that the hydrological year avoids splitting in the middle of a low flow period. In this case $P(X)$ refers to the probability of an annual minimum greater than or equal to the value X. The formulae used are the same as for flood frequency analysis (Weibull etc.).

There is one major difference between flood frequency and low flow frequency analysis which has huge implications for the statistical methods used: there is a finite limit on how low a flow can be. In theory a flood can be of infinite size, whereas it is not possible for a low flow to be less than zero (negative flows should not exist in fresh water hydrology). This places a limit on the shape of a probability distribution, effectively truncating it on the left-hand side (see Figure 6.15).

The statistical techniques described on pp. 111–112 (for flood frequency analysis) assume a full log-normal distribution and cannot be easily applied for low flows. Another way of looking at this problem is shown in Figure 6.16 where the probabilities calculated from the Weibull formula are plotted against the annual minimum flow values. The data fit a straight line, but if we extrapolate the line further it would intersect the x-axis at a value of approximately 0.95. The implication from this is that there is a 5 per cent chance of having

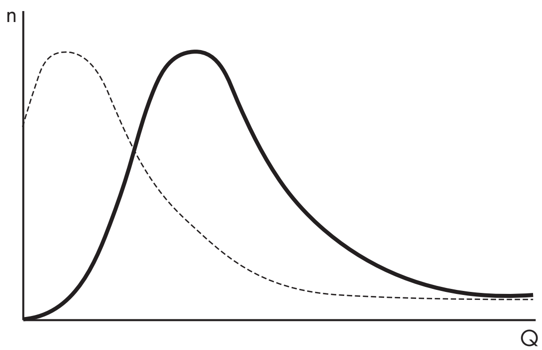

Figure 6.15 Two probability density functions. The usual log-normal distribution (solid line) is contrasted with the truncated log-normal distribution (broken line) that is possible with low flows (where the minimum flow can equal zero).

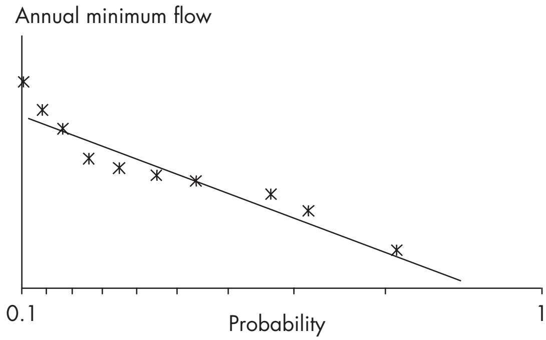

Figure 6.16 Probability values (calculated from the Weibull sorting formula) plotted on a log scale against values of annual minimum flow (hypothetical values).

a flow less than zero (i.e. a negative flow). The way around this is to fit an exponential rather than a straight line to the data. This is easy to do by eye but complicated mathematically. It is beyond the level of this text to describe the technique here (see Shaw, 1994, or Wang and Singh, 1995 for more detail). Gordon *et al*. (1992) provide a simple method of overcoming this problem, without using complicated line-fitting procedures.

Limitations of frequency analysis

As with any estimation technique there are several limitations in the application of frequency analysis; three of these are major:

1 The estimation technique is only as good as the streamflow records that it is derived from. Where the records are short or of dubious quality very little of worth can be achieved through frequency analysis. As a general rule of thumb you should not extrapolate average recurrence intervals beyond twice the length of your data set. There is a particular problem with flood frequency analysis in that the very large floods can create problems for flow gauges and therefore this extreme data may be of dubious quality (see pp.89).

2 The assumption is made that each storm or low flow event is independent of another used in the data set. This is relatively easy to guard against in annual maximum (or minimum) series, but more difficult for a peak threshold series.
3 There is an inherent assumption made that the hydrological regime has remained static during the complete period of record. This may not be true where land use, or climate change, has occurred in the catchment (see Chapter 8).

COMPUTER MODELLING IN HYDROLOGY

The easiest way of thinking about a hydrological model is to envisage streamflow as a series of numbers. Each number represents the volume of water that has flowed down the stream during a certain time period. A numerical model attempts to produce its own set of numbers, 'simulating' the flow of water down the river. There are many different ways of achieving this simulation, as will be discussed in the following section.

A model (whether mathematical, numerical or scale) is a simplification of reality. We simplify reality because the complexity of the natural world makes it difficult to understand all the processes and interactions occurring. A laboratory experiment is a similar simplification of reality; normally we are controlling all the inputs for an experiment and allowing some controlled change in a variable in order to observe the result. In constructing a computer model we are normally trying to build as good a representation of hydrological reality as we can, given our understanding of the key hydrological processes and our ability to represent these as a series of equations.

Computer modelling strategies

Black box models

The simplest forms of numerical models simulate streamflow as a direct relationship between it and another measured variable. As an example a relationship can be derived between annual rainfall and annual runoff for a catchment (see Figure 6.17). The regression line drawn to correlate rainfall and runoff is a simulation model. If you know the annual rainfall for the catchment then you can simulate the annual runoff, using the regression relationship. This type of model is referred to as a black box model as it puts all the different hydrological processes that we know influence the way that water moves from rainfall to runoff into a single regression relationship. The simplicity of this type of model makes it widely applicable but its usefulness is restricted by the end-product from the model. In the example given, the regression model may be useful to estimate annual runoff in areas with the same geology and land use but it will not tell you anything about runoff at time-scales less than one year or under different climatic and geomorphologic conditions. Another, frequently used example of a black box model is the unit hydrograph (described earlier in this chapter).

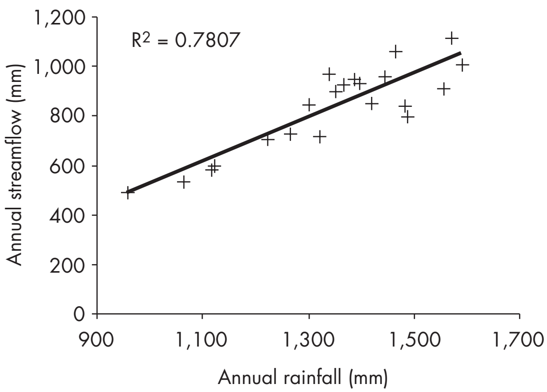

Figure 6.17 Annual rainfall vs. runoff data (1980–2000) for the Glendhu tussock catchment in the South Island of New Zealand.

Source: Data courtesy of Barry Fahey

Case study

SOIL CURVE NUMBERS FOR RAINFALL–RUNOFF RELATIONSHIP

An empirical, black-box approach to predicting runoff from rainfall is the Curve Number (CN) approach developed by the United States Department of Agriculture, Soil Conservation Service (SCS, 1972). The CN methodology has been used extensively in the USA for modelling rainfall–runoff relationships. The fundamental equation at the heart of the CN method is described in equation 6.9.

$$Q = \frac{(P - I_a)^2}{P - I_a + S} \tag{6.9}$$

where Q is the surface runoff (mm); P is the storm precipitation total (mm); I_a is the initial abstractions (all losses before runoff begins, e.g. surface storage, rainfall interception) (mm); and S is the so-called retention parameter (mm) defined in equation 6.10:

$$S = 25.4\frac{1000}{CN} - 10 \tag{6.10}$$

where CN refers to the curve number, which is derived using lookup tables (see SCS, 1986). The CN values vary according to soil type, land use, slope and changes in antecedent soil water content. The actual number of the curve is representative of the percentage of storm rainfall that runs off as stormflow, i.e. CN of 100 corresponds to all rainfall occurring as stormflow, such as for an impervious pavement (Figure 6.18).

Empirical studies on small agricultural watersheds in the USA suggest that the initial abstraction term (I_a) can be approximated using equation 6.11:

$$I_a = 0.2S \tag{6.11}$$

This reduces equation 6.9 to the form shown in equation 6.12:

$$Q = \frac{(P - 0.2S)^2}{P + 0.8S} \tag{6.12}$$

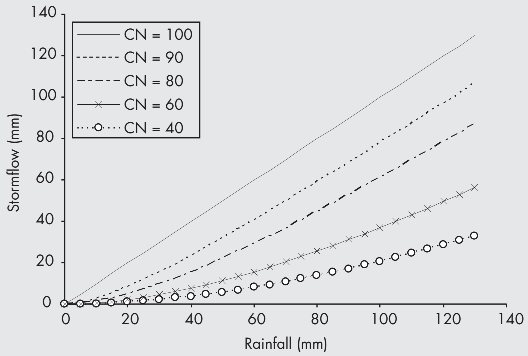

Figure 6.18 Runoff curves for a range of rainfalls.

In the CN methodology antecedent soil moisture condition is accounted for by having three different CNs, for dry, average and wet conditions.

The CN method provides a simple solution to the problem of how to model the rainfall–runoff relationship. There are other methods to model the runoff from rainfall, e.g. the modified Green-Ampt infiltration method is frequently used in physically-based hydrological models to provide infiltration and surface runoff estimates. The simplicity provided by the CN method has many attractions but it does suffer from consequent drawbacks. The most notable of these for any analysis of land use change is that CN varies according to soil characteristics and land cover. A land use change from pasture (or suburban garden) to forest will lead to an alteration in three factors: the soil infiltration characteristics; the rainfall interception; and the antecedent soil moisture conditions. Therefore more than one factor is likely to be altered in equations 6.9–6.12 and a simple alteration of CN may not be enough to properly account for the land use change. To fully account for this type of land use change it is necessary to use a hydrological model that explicitly accounts for antecedent soil moisture, soil infiltration characteristics and rainfall interception as distinct hydrological processes.

The CN approach has been used extensively to make runoff predictions based on a time series of rainfall (Ponce and Hawkins, 1996). It has also been incorporated into more sophisticated models such as the Soil and Water Assessment Tool (SWAT; Cao *et al*., 2006). In this case the rainfall–runoff relationship is derived from the CN approach and combined with other hydrological processes such as evaporation estimation and river flow rates.

Lumped conceptual models

Lumped conceptual models were the first attempt to reproduce the different hydrological processes within a catchment in a numerical form. Rainfall is added to the catchment and a water budget approach used to track the losses (e.g. evaporation) and movements of water (e.g. to and from soil water storage) within the catchment area. There are many examples in the literature of lumped conceptual models used to predict streamflows (e.g. Brandt *et al*., 1988).

The term 'lumped' is used because all of the processes operate at one spatial scale – that is, they are lumped together and there is no spatial discretisation. The scale chosen is often a catchment or sometimes sub-catchments.

The term 'conceptual' is used because the equations governing flow rates are often deemed to be conceptually similar to the physical processes operating. So, for instance, the storage of water in a canopy or the soil may be thought of as similar to storage within a bucket. As water enters the bucket it fills up until it overflows water at a rate equal to the entry rate. At the same time it is possible to have a 'hole' in the bucket that allows flow out at a rate dependent on the level of water within the bucket (faster with more water). This is analogous to soil water or canopy flow but is not a detailed description such as the Darcy–Richards approximation or the Rutter model. The rate of flow through the catchment, and hence the estimated streamflow, is controlled by a series of parameters that need to be calibrated for a given catchment. Calibration is normally carried out by comparing predicted flows to measured values and adjusting (or 'optimising') the parameters until the best fit is obtained. There is considerable debate on this technique as it may sometimes be possible to obtain a similar predicted hydrograph using a completely different set of optimised parameters. It is certainly true that the optimised parameters cannot be treated as having any physical meaning and should not be transferred to catchments other than those used for calibration.

Lumped conceptual models offer a method of formulating the hydrological cycle into a water budget model that allows simulation of streamflow while also being able to 'see' the individual processes operating. This is an advance beyond black box modelling, but because the processes are represented conceptually they are sometimes referred to as grey box models (i.e. you can see partially into them).

Physically based distributed models

The rapid advances in computing power that have occurred since the 1970s mean that numerical modelling has become much easier. Freeze and Harlan (1969) were the first to formulate the idea of a numerical model that operates as a series of differential equations in a spatially distributed sense, an idea that prior to computers was unworkable. Their ideas (with some modifications from more recent research) were put into practice by several different organisations to make a physically based distributed hydrological model. Perhaps the best known of these is the Système Hydrologique Européen (SHE) model, built by a consortium of

French, Danish and British organisations during the 1970s and early 1980s (Abbott *et al*., 1986). A model such as the SHE uses many of the process estimation techniques described in earlier chapters (e.g. Darcy's law for subsurface flow, Rutter's model for canopy interception, snow melt routines, etc.) in a water budgeting framework. Each of the equations or models used are solved for individual points within a catchment, using a grid pattern.

The principle behind this type of model is that it is totally transparent; all processes operating within a catchment are simulated as a series of physical equations at points distributed throughout the catchment. In theory this should mean that no calibration of the model is required and spatially distributed model output for any parameter can be obtained. In reality this is far from the case. There are numerous problems associated with using a physically based, distributed model, as outlined by Beven (1989), Grayson *et al*. (1992) and others. The principal problem is that the amount of data required to set the initial conditions and parameterise the model is vast. The idea of obtaining saturated hydraulic conductivity measurements for every grid point in a catchment is impossible, let alone all the other parameters required. The lack of data to run the model leads to spatial averaging of parameters. There are also concerns with the size of grid used in applications (sometimes up to 1 km^2) and whether it is feasible to use the governing equations at this scale. These types of problems led Beven (1989) to query whether there really is such a thing as physically based distributed hydrological models or whether they are really just lumped conceptual models with fancier equations.

The concept of physically based distributed hydrological modelling is noble, but in reality the models have not produced the results that might have been expected. They are certainly unwieldy to use and have many simplifications that make the terminology doubtful. However they have been useful for gaining a greater understanding of our knowledge base in hydrological processes. The approach taken, with its lack of reliance on calibration, still offers the only way of investigating issues of land use change and predicting flows in ungauged catchments.

Hydrological modelling for specific needs

In many cases where streamflow needs to be estimated, the use of a physically based model is akin to using the proverbial 'sledgehammer to crack a walnut'. With the continuing increase in computing power there are numerous tools available to the hydrologist to build their own computer model to simulate a particular situation of interest. These tools range from Geographic Information Systems (GIS) with attached dynamic modelling languages to object-oriented languages that can use icon-linked modelling approaches (e.g. McKim *et al*., 1993). This perhaps offers a future role for hydrological modelling away from the large modelling packages such as SHE. In essence it allows the hydrologist to simulate streamflow based on a detailed knowledge of catchment processes of importance for the particular region of interest.

FLOW ASSESSMENT FOR STREAM ECOLOGY

Managers of river systems frequently need information on the amount of flow required to sustain the current river ecology. This is in order to ascertain how much water is available for out of stream usage (e.g. irrigation) without placing detrimental stress on the current river ecology. Consequently a branch of science has been developed that combines knowledge of river hydraulics with aquatic ecology to provide this information. This is described in the following section. The discussion is continued further in Chapter 8 where there is a description of water allocation methodology.

Jowett (1997) divides the methodology used for instream flow assessment into three types with an increasing complexity: *historic flow*, *hydraulic* and *habitat methods*. The historic approach sets water abstraction limits based on a historical flow range; the hydraulic method uses the relationship between

Case study

WATYIELD – MODELLING CHANGES IN WATER YIELD FROM ALTERING LAND COVER CHANGE

The degree of vegetation cover in a catchment will affect the amount of water flowing down a stream. The physical processes that cause this effect have been described in Chapter 3 and the impacts of this change are discussed in Chapter 8. A water balance model has been developed to quantify the impact of land use changes on the stream discharge. The model is simple to use and can be downloaded for free from the World Wide Web (look for WATYIELD at http://icm.landcare research.co.nz). Also available at this site are a series of reports that help parameterise the model and a full user's guide.

The WATYIELD model was developed for New Zealand conditions and works best in a humid temperate environment. It has been designed for catchments up to around 50km^2 in size. In the modelling terminology outlined earlier in this chapter, WATYIELD could be described as a lumped, conceptual model. However, there is detailed process representation of rainfall interception and soil moisture storage within the model so it moves slightly towards being physically based. The spatial representation is at the catchment scale; although it is possible to split a catchment into sub-sections with different vegetation covers and rainfall distributions. However, these sub-sections have no spatial differentiation within the catchment, i.e. the model doesn't know where they are within a catchment, just that there are subsections. For catchments larger than 50km^2 the underlying assumptions of spatial uniformity start to breakdown and it is necessary to start introducing elements such as flow routing down a stream (presently ignored at the daily time step of WATYIELD).

WATYIELD works by adding daily rainfall to two storage terms which release water to a river based on hydrograph recession coefficients (a full description of the model can be found in Fahey *et al*., 2004). The storage terms represent soil moisture and a deeper groundwater store. Daily rainfall is processed by the model so that any interception loss from a vegetation canopy is removed and all the resultant rainfall infiltrates into the soil moisture store. In order to operate the model a daily rainfall record, potential evapotranspiration, soil parameters and knowledge about flow characteristics from a nearby stream are required. Much of this type of data is readily available from the scientific literature and resource management databases.

WATYIELD has been applied to a 23 km^2 catchment (Rocky Gully) in the South Island of New Zealand to investigate two possible land use change scenarios. The current vegetation cover for the catchment is a mixture of tall tussock grassland, pasture grasses and a small amount of scrubland forest. The catchment has an altitude range from 580 m to 1,350 m with an increasing rainfall with altitude.

In testing the model against daily streamflow from 1989–2001, WATYIELD was able to predict mean annual flow within 2 per cent accuracy and mean annual seven-day low flow within 3 per cent. The two scenarios simulated were:

- 40 per cent of the catchment was converted to plantation forestry (*Pinus radiata*). All the planting occurred in the lower half of the catchment;
- 50 per cent of the catchment was converted from tussock to pasture grassland. All the tussock grassland in the upper half of the catchment was replaced with pasture species.

Each of these is a realistic land use change scenario for the region; conversion of pasture land to forestry is common practice, as is 'improving' grassland by over-sowing with rye grass species. The land use change scenarios were simulated in the model using the 1989–2001 rainfall data (i.e. repeating the earlier simulation but with a different land cover).

The results from the modelling are shown in Table 6.5. An initial look at the results suggests a surprising result: the amount of interception loss from a 40 per cent increase in forestry does not transfer through into much of a change in mean annual streamflow or low flows. There is a larger change in flow regime from the replacement of tussock grassland in the upper catchment; despite this land use change resulting in a lowering of interception loss (tussock grassland has higher interception losses than pasture grass). The reason for these results is that it is the upper part of the catchment, with a higher rainfall, that produces most of the streamflow, particularly the low flows. Hence a change in land use in the lower section makes a relatively small change in the flow regime. However, a change in land use in the upper region of the catchment has a larger effect because this is where the effective rainfall is occurring. A change from tussock to pasture grassland increases the transpiration loss which more than offsets the decrease in canopy interception.

In this case WATYIELD was able to tease out the difference between canopy interception and canopy transpiration. The difference between the two is what made the most difference in the simulations. The final scenario modelled was to place the forestry in the upper reaches of the catchment; this reduced flows by around 25 per cent. However this is a highly unlikely land use change scenario as commercial forestry at this latitude does not normally extend beyond 850 m above sea level.

Table 6.5 Results from WATYIELD modelling of land use change

Flow measure	*Scenario 1 (forestry in lower half replaces pasture)*	*Scenario 2 (replacement of tussock grassland with pasture)*
Mean annual flow	Reduced by 6%	Reduced by 7%
Mean annual 7-day low flow	Reduced by 3%	Reduced by 7%

hydraulic parameters (e.g. wetted perimeter, stream depth, etc.) and stream health; and the habitat method uses actual measurements of stream health with changes in flow regime to predict the impacts of flow changes. The way these methods treat the relationship between increased streamflow and the biological response is shown in Figure 6.19. The historic method assumes that there is a linear relationship so that more flow results in a greater biological response. The hydraulic method recognises that stream beds are non-linear in form and therefore a small change in flow may result in large increases in biological productivity but that this decreases as the flow increases. The habitat method recognises that there is a maxima in the biological productivity and high flows may lead to decreasing biological response.

The most well known of the *historic flow methods* is 'Montana method' proposed by Tennant (1976), also called the Tennant method. Tennant (1976) used hydraulic data from eleven streams in the USA and knowledge about depths and velocities required to sustain aquatic life to suggest that 10 per cent of average flow is the lower limit for aquatic life. Tennant (1976) also recommended that 30 per cent of average flow provides a satisfactory stream

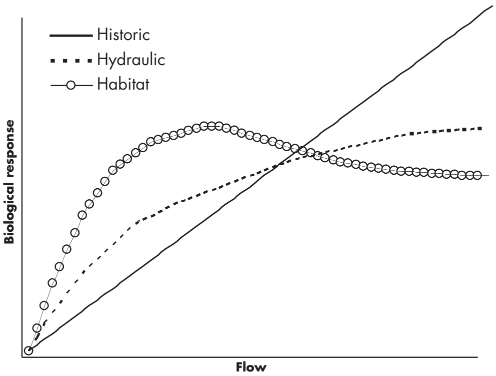

Figure 6.19 Hypothetical relationships showing biological response to increasing streamflow as modelled by historic, hydraulic and habitat methods.

Source: Adapted from Jowett (1997)

environment. With relatively easily derived flow information (i.e. average flow) a new flow regime can be set for a river that takes into account the instream values. However this approach precludes the possibility that a stream could be enhanced by a non-natural flow regime. This is especially true where there is an upstream reservoir, in which case flows can be manipulated to improve the aquatic environment, not just maintain what is presently there.

The *hydraulic method* requires measurements of hydraulic data such as wetted perimeter, width, velocity and depth at a series of cross sections. Then, using either rating curves (i.e. the stage–discharge relationship described in Chapter 5) or an equation such as Manning's (see Chapter 5), the variations in a hydraulic parameter with flow can be derived. The most commonly used hydraulic parameter is the wetted perimeter because it takes into account the area of streambed where periphyton and invertebrates live. A healthy periphyton and invertebrate community generally leads to a healthy river ecosystem. The variation in wetted perimeter with flow is drawn in the same way as represented by the broken line in Figure 6.19. The minimum flow for river is normally defined by where the hydraulic parameter (e.g. wetted perimeter) starts to decline sharply to zero. The hydraulic method has the advantage over the historic method that it takes into account the actual streambed morphology which may differ markedly between rivers.

The *habitat method* extends the hydraulic method by taking the hydraulic information and combining it with knowledge of how different aquatic species survive in those flow regimes. In this way the appropriate flow regime can be designed with particular aquatic species in mind. In the case of fish, some prefer shallow turbulent streams compared to deep, slow moving rivers. The habitat method allows differentiation between these so that a flow regime can be set with protection, or enhancement, of a particular species in mind.

The most common use of the habitat method is the Instream Flow Incremental Methodology (IFIM; Irvine *et al*., 1987; Navarro *et al*., 1994) which has been developed into computer models such as PHABSIM (Physical HABitat SIMulation; Milhous *et al*., 1989; Gallagher and Gard, 1999) and RHYHABSIM (River HYdraulic HABitat SIMulation; Jowett, 1997).

The habitat method focuses on a particular species and life stage at a time, and investigates its response at a particular flow. For each cell in a two-dimensional grid, velocity, depth, substrate and possibly other parameters (e.g. cover) at the given flow are converted into suitability values, one for each parameter. These suitability values are combined (usually multiplied) and multiplied by the cell area to give an area of usable habitat (also called weighted usable area, WUA). Finally, all the usable habitat cell areas are summed to give a total habitat area (total WUA) for the reach at the given flow. The whole procedure is repeated for other flows until a graph of usable habitat area versus flow for the given species has been produced. This graph has a typical shape, as shown in Figure 6.19, with a rising part, a maximum and a decline. The decline occurs when the velocity and/or depth exceed those preferred by the given species and life stage. In large rivers, the curve may predict that physical habitat will be at a maximum at less than naturally occurring flows (Jowett, 1997).

Models such as PHABSIM have been used successfully in many places around the world to advise water managers on the best flow regime for particular aquatic species. It should be noted that this is a physical approach to the problem, i.e. it takes account of the physical flow regime of the river with no consideration of water quality parameters. The assumption is made that water quality does not change with the proposed changing flow regime.

SUMMARY

The analysis of streamflow records is extremely important in order to characterise the flow regime for a particular river. Hydrograph analysis involves dissecting a hydrograph to distinguish between stormflow and baseflow. This is often a precursor to using the unit hydrograph, a technique using past stormflow records to make predictions on the likely form of future storm events. Flow duration curves are used to look at the overall hydrology of a river – the percentage of time a river has an average flow above or below a certain threshold. Frequency analysis is used to look at the average return period of a rare event (or the probability of a certain rare event occurring), whether that be extremes of flooding or low flow. Each of the methods described in this chapter has a distinct use in hydrology and it is important that practising hydrologists are aware of their role.

Computer modelling offers a methodology to investigate hydrological processes and make predictions on what the flow might be in a river given a certain amount of rainfall. There are different types of models, with differing amounts of complexity, but all are a simplification of reality and aim to either make a prediction or improve our understanding of biophysical processes.

A key use of flow and hydraulic data is to assess the needs of aquatic fauna within a river system. This information is used by managers of regulated rivers to set flow regimes that are beneficial, or at least non-detrimental, to particular flow species. Models such as PHABSIM have been used successfully in many places around the world to achieve this. They combine hydrological and ecological knowledge to provide vital information to resource managers.

ESSAY QUESTIONS

1. **Find a scientific paper in the literature that uses a hydrological model and evaluate the type of model and its strengths and weaknesses for the study concerned.**
2. **Outline the limitations of the unit hydrograph when used as a predictive tool and attempt to explain its success despite these limitations.**
3. **Describe the types of information that can be derived from a flow duration curve and explain the use of that information in hydrology.**
4. **Explain why interpretation of flood (or low flow) frequency analysis may be fraught with difficulty.**
5. **Describe the data (and measurement equipment) required for using the IFIM approach to look at the habitat requirements for a particular aquatic species.**

FURTHER READING

Beven, K.J. (2001) *Rainfall-runoff modelling: the primer*. Wiley, Chichester.
An introduction to modelling in hydrology.

Callow, P. and Petts G.E. (eds.) (1994) *The rivers handbook: hydrological and ecological principles*. Blackwell Publishing.
Gives further detail on combining hydraulic, hydrological and ecological principles for river management.

Dingman, S.L. (1994) *Physical hydrology*. Macmillan, New York.
A high level text with good detail on analytical techniques.

Maidment, D.R. (ed.) (1993) *Handbook of hydrology*. McGraw-Hill, New York.
Detailed text with Part 3 concentrating on hydrological analysis techniques.

Shaw, E.M. (1994) *Hydrology in practice* (3rd edition). Chapman and Hall, London.
An engineering text with good detail on analytical techniques.

7

WATER QUALITY

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of water quality as an issue in hydrology and how it ties into water quantity.
- A knowledge of the main parameters used to assess water quality and what affects their levels in a river.
- A knowledge of the measurement techniques and sampling methodology for assessing water quality.
- A knowledge of techniques used to control water pollution and manage water quality.

This chapter identifies the different types of pollutants that can be found in a river system and describes the major sources of them, especially where elevated levels may be found and what impact their presence has on aquatic ecology. The chapter also outlines the methods used to measure water quality parameters. This is followed with a description of the management techniques used to control water quality in a river catchment.

Traditionally hydrology has been interested purely in the amount of water in a particular area: water quantity. This is frequently referred to as physical hydrology. If, however, we take a wider remit for hydrology – to include the availability of water for human consumption – then issues of water quality are of equal importance to quantity. There are three strong arguments as to why hydrology should consider water quality an area worthy of study.

1 *The interlink between water quality and quantity*. Many water quality issues are directly linked to the amount of water available for dilution and dispersion of pollutants, whether they be natural or anthropogenic in source. It is virtually impossible to study one without the other. An example of this is shown in the Case Study of the River Thames through London (pp. 127–129).
2 *The interlink between hydrological processes and water quality*. The method by which pollutants transfer

from the land into the aquatic environment is intrinsically linked with the hydrological pathway (i.e. the route by which the water moves from precipitation into a stream), and hence the hydrological processes occurring. A good example of this is in Heppell *et al*. (1999) where the mechanisms of herbicide transport from field to stream are linked to runoff pathways in a clay catchment.

3 *Employment of hydrologists*. It is rare for someone employed in water resource management to be entirely concerned with water quantity, with no regard for quality issues. The maintenance of water quality is not just for drinking water (traditionally an engineer's role) but at a wider scale can be for maintaining the **amenity value** of rivers and streams.

It is easy to think of water quality purely in terms of pollution; i.e. waste substances entering a river system as a result of human activity. This is an important issue in water-quality analysis but is by no means the only one. One of the largest water quality issues is the amount of suspended sediment in a river, which is frequently a completely natural process. Suspended sediment has severe implications for the drinking-water quality of a river, but also for other hydrological concerns such as reservoir design and aquatic flora and fauna. As soon as a river is dammed the water velocity will slow down. Simple knowledge of the **Hjulstrom curve** (see Figure 7.1) tells us that this will result in the deposition of suspended sediment. That deposition will eventually reduce the capacity of the reservoir held behind the dam. In high-energy river systems, for even a very large reservoir, a dramatic reduction in capacity can take place within two to four decades. It is critically important for a hydrologist involved in reservoir design to have some feeling for the quantities of suspended sediment so that the lifespan of a reservoir can be calculated. In South Korea, reservoir management includes understanding the sediment plume entering a reservoir during the rainy season and using a multiple level abstraction to release this sediment laden water during the wet season, i.e. avoiding sedimentation in the dam (Kim *et al*., 2007).

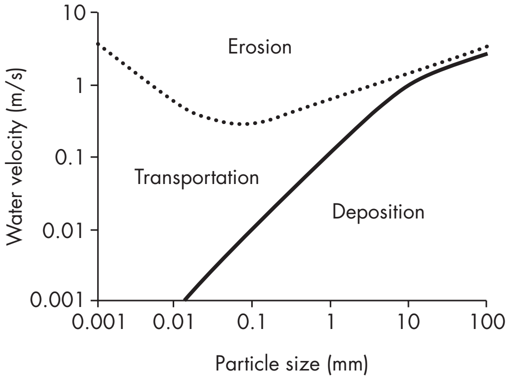

Figure 7.1 The Hjulstrom curve relating stream velocity to the erosion/deposition characteristics for different sized particles (x-axis). In general the slower the water moves the finer the particles that are deposited and the faster the water moves the larger the particles being transported.

Spatial variations in water quality may be influenced by many different environmental factors (e.g. climate, geology, weathering processes, vegetation cover and anthropogenic). Often it is a combination of these factors that makes a particular water-quality issue salient for a particular area. An example of this is acid rain (also discussed in Chapter 2) as a particular problem for north-eastern North America and Scandinavia. The sources of the acid rain are fossil-fuel-burning power stations and industry. It becomes a particular problem in these areas for a number of reasons: it is close to the sources of acid rain; high rainfall contributes a lot of acid to the soil; the soils are thin after extensive glaciation and derived from very old rocks; and the soils are heavily leached (have had a lot of water passing through them over a long time period) and have a low buffering capacity (see p. 34). This combination of influences means that the water in the rivers has a low **pH**, and – of particular concern to gill-bearing aquatic fauna – has a high dissolved aluminium content (from the soils).

Having argued for the role of natural water-quality issues to be considered seriously, the reader

Case study

THE RIVER THAMES THROUGH LONDON: WATER QUALITY CHANGE

The River Thames as it flows through London is one of the great tourist sights of Europe. It is an integral part of London, not just for its scenic attraction but also as a transport route right into the heart of a modern thriving city. The river has also a large part to play in London's water resources, both as source of drinking water and a disposal site for waste.

London has a long history of water-quality problems on the Thames, but it has not always been so. Prior to the nineteenth century domestic waste from London was collected in cesspools and then used as fertiliser on agricultural land (hence the use of the term 'sewage farm' for sewage treatment stations). The Thames maintained a fish population, and salmon from the river were sold for general consumption. With the introduction of compulsory water closets (i.e. flushing toilets) in 1843 and the rise in factory waste during the Industrial Revolution, things started to change dramatically for the worse during the nineteenth century. The majority of London's waste went through poorly constructed sewers (often leaking into shallow aquifers which supplied drinking water) straight into the Thames without any form of treatment. In 1854 there was an outbreak of cholera in London that resulted in up to 10,000 deaths. In a famous epidemiological study Dr John Snow was able to link the cholera to sewage pollution in water drawn from shallow aquifers. The culmination of this was 'the year of the great stink' in 1856. The smell of untreated waste in the Thames was so bad that disinfected sheets had to be hung from windows in the Houses of Parliament to lessen discomfort for the lawmakers of the day. In the best NIMBY ('not in my backyard') tradition this spurred parliament into action and in the following decade, radical changes were made to the way that London used the River Thames. Water abstraction for drinking was only permitted upstream of tidal limits and London's sewage was piped downstream to Beckton where it was discharged (still untreated) into the Thames on an ebb tide.

The result of these reforms was a radical improvement of the river water quality through central London; but there was still a major problem downstream of Beckton. The improvements were not to last, however, as by the middle of the twentieth century the Thames was effectively a dead river (i.e. sustained no fish population and had a dissolved oxygen concentration of zero for long periods during the summer). This was the result of several factors: a rapidly increasing population, increasing industrialisation, a lack of investment in sewage treatment and bomb damage during the Second World War.

Since the 1950s the Thames has been steadily improving. Now there is a resident fish population and migratory salmon can move up the Thames. This improvement has been achieved through an upgrading of the many sewage treatment works that discharge into the Thames and its tributaries. The England and Wales Environment Agency has much to do with the management of the lower Thames and proudly proclaims that the Thames 'is one of the cleanest metropolitan rivers in the world'. How realistic is this claim?

There is no doubt that the Thames has been transformed remarkably from the 'dead' river of sixty years ago into something far cleaner, but there are two problems remaining for the management of the Thames through London, and for one of these nothing can be done.

- The Thames is a relatively small river that does not have the flushing potential of other large rivers; therefore it cannot cleanse itself very easily.

- The sewer network underneath London has not been designed for a large modern city and cannot cope with the strains put on it.

At Westminster (in front of the Houses of Parliament) the Thames is over 300 m wide; this is confined from the width of 800 m evident during Roman times. This great width belies a relatively small flow of fresh water. It appears much larger than in reality because of its use for navigation and the tidal influence. The average flow rate for the Thames is 53 cumecs, rising to 130 cumecs under high flows. In Table 7.1 this is compared with rivers that flow through other major cities. In Seoul, a similar sized capital city, the Han River is over seven times larger than the Thames. The effect of the small flow in the Thames is that it does not have great flushing power. During the summer months it may take a body of water three months to move from west London to the open sea. On each tide it may move up to 14 km in total but this results in less than a kilometre movement downstream. If this body of water is polluted in some way then it is not receiving much dilution or dispersion during the long trip through London.

The second important factor is the poor state of London's sewers. Prior to Sir Joseph Bazalgette's sewer network of 1864 the old tributaries of the Thames acted as sewers, taking waste water directly to the Thames. Bazalgette's grand sewerage scheme intercepted these rivers and transported the sewage through a large pipe to east London. This system still exists today. The actual sewerage network is very well built and still works effectively. The problem is that it is unable to cope with the volume of waste expected to travel through it, particularly when it rains, as it is a combined stormwater–sewage network. The original tributaries of the Thames, such as the Fleet, still exist under London and any storm runoff is channelled into them. When the volume of stormwater and sewage is too great for the sewers the rivers act as overspills and take the untreated sewage directly into the Thames. This is a particular problem during summer storms when the volume of water flowing down the Thames is low and cannot dilute the waste effectively. To combat this problem Thames Water Utilities (part of the private company that treats London's sewage) operate two boats especially designed to inject oxygen directly into the water. These boats can float with a body of sewage-polluted water, injecting oxygen so that the dissolved oxygen level does not reach levels that would be harmful to fish and other aquatic creatures. To help in the tracking of a polluted body of water there are water-quality monitoring stations attached to bridges over the Thames. These stations measure temperature, dissolved oxygen concentration and electrical

Table 7.1 Comparison of rivers flowing through major cities

River	*Mean annual flow (m^3/s)*	*City on river or estuary*	*Population in metropolitan area (million)*
Thames	82	London	12.0
Seine	268	Paris	9.93
Hudson	387	New York	19.3
Han	615	Seoul	10.3
Rhine	2,219	Rotterdam	1.1
Paraná/Uruguay	22,000	Buenos Aires	11.6

Source: Flow data from Global Runoff Data Centre conductivity at fifteen-minute intervals and are monitored by the Environment Agency as they are received in real time at the London office.

In addition to the oxygen-injecting boats, there is tight water-quality management for the River Thames through London. This is operated by the Thames Estuary Partnership, a group of interested bodies including the Environment Agency. Their remit includes other factors such as protecting London from flooding (using the Thames Barrier), but also setting higher effluent standards for sewage treatment works during the summer. The emphasis is on flexibility in their management of the Thames. There is no question that the River Thames has improved from fifty years ago. In many respects it is a river transformed, but it still has major water-quality problems such as you would expect to find where a small river is the receptacle for the treated waste of over 10 million people. The water-quality management of a river like the Thames needs consideration of many facets of hydrology: understanding pollutants, knowledge of stormflow peaks from large rainfall events, and streamflow statistics.

will find that the majority of this chapter deals with human-induced water-quality issues. This is an inevitable response to the world we live in where we place huge pressures on the river systems as repositories of waste products. It is also important to study these issues because they are something that humans can have some control over, unlike many natural water-quality issues.

Before looking at the water-quality issues of substances within a river system it is worth considering how they reach a river system. In studying water pollution it is traditional to differentiate between *point source* and *diffuse* pollutants. As the terminology suggests, point sources are discrete places in space (e.g. a sewage treatment works) where pollutants originate. Diffuse sources are spread over a much greater land area and the exact locations cannot be specified. Examples of diffuse pollution are excess fertilisers and pesticides from agricultural production. The splitting of pollutants into diffuse and point sources has some merit for designing preventative strategies but like most categorisations there are considerable overlaps. Although a sewage treatment works can be thought of as a point source when it discharges effluent into a stream, it has actually gathered its sewage from a large diffuse area. If there is a particular problem with a sewage treatment works effluent, it may be a result of accumulated diffuse source pollution rather than the actual sewage treatment works itself.

A more useful categorisation of water pollutants is to look at their impacts on the river system. In this way we can differentiate between three major types of pollutants.

- *Toxic compounds*, which cause damage to biological activity in the aquatic environment.
- *Oxygen balance affecting compounds*, which either consume oxygen or inhibit the transfer of oxygen between air and water. This would also include thermal pollution as warm water does not hold as much dissolved oxygen as cold water (see p. 134).
- *Suspended solids* – inert solid particles suspended in the water.

Whether we approve or not, rivers are receptacles for large amounts of waste produced by humans. Frequently this is deliberate and is due to the ability of rivers to cope with waste through degradation, dilution and dispersion. Just how quickly these three processes operate is dependent on the pollutant load already present in the river, the temperature and pH of the water, the amount of water flowing down the river and the mixing potential of the river. The last two of these are river flow characteristics that will in turn be influenced by the time of year, the nature of flow in the river (e.g. the shape of the flow duration curve), and the velocity and turbulence of flow. This demonstrates the strong interrelationship that exists between water quality and water quantity in a river system.

One remarkable feature about rivers is that given enough time and a reasonable pollution loading, rivers will recover from the input of many pollutant types. That is not to say that considerable harm cannot be done through water pollution incidents, but by and large the river system will recover so long as the pollution loading is temporary. An example of this can be seen in the **oxygen sag curve** (see Figure 7.2) that is commonly seen below point sources of organic pollution (e.g. sewage effluent). The curve shows that upon entering the river there is an instant drop in dissolved oxygen content. This is caused by bacteria and other micro-organisms in the river feeding on the organic matter in the stream and using any available dissolved oxygen. This would have a severe impact on any aquatic fauna unable to move away from this zone of low dissolved oxygen. As the pollutant load moves downstream the degradation, dilution and dispersal starts to take effect and oxygen levels start to recover in the river. The shape of the curve, especially the distance downstream until recovery, is highly dependent on the flow regime of the receiving river. A fast flowing, readily oxygenated stream will recover much faster than a slow-moving river. Large rivers will have a faster recovery time (and the depth of sag will be less) than small streams, due to the amount of dilution occurring.

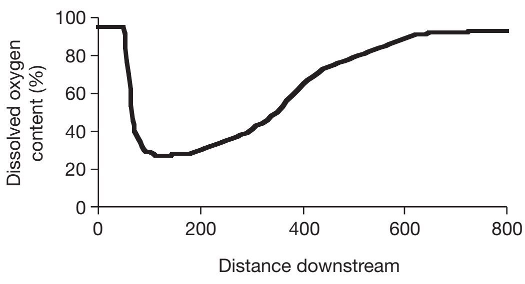

Figure 7.2 Hypothetical dissolved oxygen sag curve. The point at which the curve first sags is the point source of an organic pollutant. The distance downstream has no units attached as it will depend on the size of the river.

WATER-QUALITY PARAMETERS

To analyse the water quality within a river, consideration has to be given to what type of test may be carried out and the sampling pattern to be used. There are numerous parameters that can be measured, and each is important for the part they play in an overall water-quality story. It is not necessary to measure them all for a single water-quality analysis study; instead the relevant parameters for a particular study should be identified. This can be done using a priori knowledge of the water-quality issues being studied. To aid in this, different parameters are discussed here with respect to their source; what type of levels might be expected in natural rivers; and the impact they have on a river ecosystem.

The first distinction that can be made is between physical and chemical parameters. With chemical parameters it is the concentration of a particular chemical substance that is being assessed. With physical parameters it is a physical measurement being made, normally measuring the amount of something within a water sample.

Physical parameters

Temperature

The temperature of water in a river is an important consideration for several reasons. The most important feature of temperature is the interdependence it has with dissolved oxygen content (see p. 134). Warm water holds less dissolved oxygen than colder water. The dissolved oxygen content is critical in allowing aquatic fauna to breathe, so temperature is also indirectly important in this manner. Water temperature is also a controlling factor in the rate of chemical reactions occurring within a river. Warm water will increase the rate of many chemical reactions occurring in a river, and it is able to dissolve more substances. This is due to a weakening of the hydrogen bonds and a greater ability of the bipolar molecules to surround anions and cations.

Warm water may enter a river as thermal pollution from power stations and other industrial processes. In many power stations (gas, coal and nuclear) water is used as a coolant in addition to the generation of steam to drive turbines. Because of this, power stations are frequently located near a river or lake to provide the water source. It is normal for the power stations to have procedures in place so that hot water is not discharged directly into a river; however, despite the cooling processes used, the water is frequently 1–2°C degrees warmer on discharge. The impact that this has on a river system will be dependent on the river size (i.e. degree of dilution and rate of dispersion).

Dissolved solids

In the first chapter, the remarkable ability of water to act as a solvent was described. As water passes through a soil column or over a soil surface it will dissolve many substances attached to the soil particles. Equally water will dissolve particles from the air as it passes through the atmosphere as rain. The amount of dissolved substances in a water sample is referred to as the **total dissolved solids (TDS)**. The higher the level of TDS the more contaminated a water body may be, whether that be from natural or anthropogenic sources. Meybeck (1981) estimates that the global average TDS load in rivers is around 100 mg/l, but it may rise considerably higher (e.g. the Colorado River has an average TDS of 703 mg/l).

Electrical conductivity

A similar measurement to TDS is provided by the electrical conductivity. The ability of a water sample to transmit electrical current (its conductivity) is directly proportional to the concentration of dissolved ions. Pure, distilled water will still conduct electricity but the more dissolved ions in water the higher its electrical conductivity. This is a straight-line relationship, so equation 7.1 can be derived.

$$K = \frac{\text{Conductivity}}{\text{TDS}} \text{ or } \text{TDS} = \frac{\text{Conductivity}}{K} \qquad (7.1)$$

This relationship gives a very good surrogate measure for TDS. The K term is a constant (usually between 0.55 and 0.75) that can be estimated by taking several measurements of conductivity with differing TDS levels. Conductivity is a simple measurement to take as there are many robust field instruments that will give an instant reading. This can then be related to the TDS level at a later stage. Electrical conductivity is measured in Siemens per metre, although the usual expression is microsiemens per centimetre (μS/cm). Rivers normally have a conductivity between 10 and 1,000 μS/cm.

Suspended solids

The amount of suspended solids has been highlighted at the start of this chapter as a key measure of water quality. The carrying of suspended sediment in a river is part of the natural erosion and sediment transport process. The sediment will be deposited at any stage when the river velocity drops and conversely it will be picked up again with higher river velocities (see Figure 7.1). In this manner the **total suspended solids (TSS)** load will vary in space and time. The amount of TSS in a river will affect the aquatic fauna, because it is difficult for egg-laying fish and invertebrates to breed in an environment of high sediment. Suspended sediment is frequently inert, as in the case of most clay and silt particles, but it can be organic in content and therefore have an oxygen demand.

TSS is expressed in mg/l for a water sample but frequently uses other units when describing sediment load. Table 7.2 shows some values of sediment discharge (annual totals) and calculates an average TSS from the data. It is remarkable to see the data in this form, enabling contrast to be drawn between the different rivers. Although the Amazon delivers a huge amount of sediment to the oceans it has a relatively low average TSS, a reflection of the extremely high discharge. In contrast to this the

Table 7.2 Sediment discharge, total river discharge (averaged over several years) and average total suspended solids (TSS) for selected large river systems

River (country)	*Sediment discharge (10^3 tonnes/yr)*	*Discharge (km^3/yr)*	*Average TSS (mg/l)*
Zaïre (Zaïre)	43,000	1,250	0.03
Amazon (Brazil)	900,000	6,300	0.14
Danube (Romania)	67,000	206	0.33
Mississippi (USA)	210,000	580	0.36
Murray (Australia)	30,000	22	1.36
Ganges-Brahmaputra (Bangladesh)	1,670,000	971	1.72
Huanghe or Yellow (China)	1,080,000	49	22.04

Source: Data from Milliman and Meade (1983)

Huanghe river (sometimes referred to as the Yellow river due to the high sediment load) is virtually a soup! It must be noted that these are average values over a year and that the TSS will vary considerably during an annual cycle (the TSS will rise considerably during a flood).

Turbidity

A similar measure to TSS is the **turbidity**: a measure of the cloudiness of water. The cloudiness is caused by suspended solids and gas bubbles within the water sample, so TSS and turbidity are directly related. Turbidity is measured as the amount of light scattered by the suspended particles in the water. A beam of light of known luminosity is shone through a sample and the amount reaching the other side is measured. This is compared to a standard solution of formazin. The units for turbidity are either FTU (formazin turbidity units) or NTU (normalised turbidity units); they are identical. Turbidity is a critical measure of water quality for the same reasons as TSS. It is a simpler measurement to make, especially in the field, and therefore it is sometimes used as a surrogate for TSS.

Chemical parameters

pH

Chemists think of water as naturally disassociating into two separate ions: the hydroxide (OH^-) and hydrogen (H^+) ions.

$$H_2O \rightleftharpoons OH^- + H^+$$

The acidity of water is given by the hydrogen ion, and hence pH (the measure of acidity) is a measure of the concentration of hydrogen ions present. In fact it is the log of the inverse concentration of hydrogen ions (equation 7.2).

$$pH = \log\frac{1}{\left[H^+\right]} \tag{7.2}$$

This works out on a scale between 1 and 14, with 7 being neutral. A pH value less than 7 indicates an acid solution; greater than 7 a basic solution (also called alkaline). It is important to bear in mind that because the pH scale is logarithmic (base 10) a solution with pH value 5 is ten times as acidic as one with pH value 6.

In natural waters the pH level may vary considerably. Rainwater will naturally have a pH value less than 7, due to the absorption of gases such as carbon dioxide by the rainwater. This forms a weak carbonic acid, increasing the concentration of hydrogen ions in solution. The normal pH of rainfall is somewhere between 5 and 6 but may drop as low as 4, particularly if there is industrial air pollution nearby. For example, Zhao and Sun (1986) report a pH value of 4.02 in Guiyang city, China, during 1982.

Acidic substances may also be absorbed easily as water passes through a soil column. A particular example of this is water derived from peat, which will absorb organic substances. These form organic acids, giving peat-derived water a brown tinge and a low pH value. At the other end of the spectrum rivers that drain carbonate-rich rocks (e.g. limestone and chalk), have a higher pH due to the dissolved bicarbonate ions.

The pH value of rivers is important for the aquatic fauna living within them. The acidity of a river is an important control for the amount of dissolved ions present, particularly metal species. The more acidic a river is the more metallic ions will be held in solution. For fish it is often the level of dissolved aluminium that is critical for their survival in low pH waters. The aluminium is derived from the breakdown of alumino-silicate minerals in clay, a process that is enhanced by acidic water. Water with a pH between 6 and 9 is unlikely to be harmful to fish. Once it drops below 6 it becomes harmful for breeding, and salmonid species (e.g. trout and salmon) cannot survive at a pH lower than 4. Equally a pH higher than 10 is toxic to most fish species (Alabaster and Lloyd, 1980). Table 7.3 summarises the effect of decreasing pH (i.e. increasing acidity) on aquatic ecology.

Mention needs to be made of the confusing terminology regarding **alkalinity**. Alkalinity is a measure of the capacity to absorb hydrogen ions without a change in pH (Viessman and Hammer, 1998). This is influenced by the concentration of hydroxide, bicarbonate or carbonate ions. In water-quality analysis the term 'alkalinity' is used almost exclusively to refer to the concentration of bicarbonate (HCO^{3-}) ions because this is the most variable of the three. The bicarbonate ions are derived from the percolation of water through calcareous rocks (e.g. limestones or chalk). It is important to know their concentration for the buffering of pH and for issues of water hardness. The buffering capacity of soils, and water derived from soils, is an important concept in water quality. The buffering capacity of a solution is the ability to absorb acid without changing the pH. This is achieved through a high base cation load or high bicarbonate load. This is why soil derived from limestone and chalk has fewer problems coping with acid rain.

Dissolved oxygen

Dissolved oxygen is vital to any aquatic fauna that use gills to breath. Salmonid species of fish require dissolved oxygen contents greater than 5 mg/l,

Table 7.3 Effect of increasing acidity on aquatic ecology

Effect on organisms or process	*pH value*
Mayflies disappear	6.5
Phytoplankton species decline – green filamentous periphyton appears	6
Molluscs disappear	5.5–6.0
Waterfowl breeding declines	5.5
Bacterial decomposition slows/fungal decomposition appears	5
Salmonid reproduction fails – aluminium toxicity increases	5
Most amphibia disappear	5
Caddis flies, stone flies and Megaloptera (dobsonflies, alderflies, etc.) disappear	4.5–5.0
Beetles, dragonflies and damselflies disappear	4.5
Most adult fish harmed	4.5

Source: Dodds (2002), adapted from Jeffries and Mills (1990)

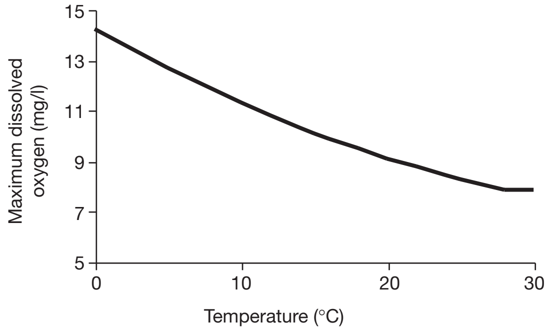

Figure 7.3 Relationship between maximum dissolved oxygen content (i.e. saturation) and temperature.

whereas coarse fish (e.g. perch, pike) can survive in levels as low as 2 mg/l. The dissolved oxygen content is also an important factor in the way we taste water. Water saturated in oxygen tastes fresh to human palates; hence drinking water is almost always oxygenated before being sent through a pipe network to consumers.

There are two methods by which dissolved oxygen content is considered: percentage saturation and concentration (mg/l). These two measures are interrelated through temperature, as the dissolved oxygen content of water is highly temperature dependent (see Figure 7.3).

Biochemical oxygen demand

One of the key water-quality parameters is the five-day biochemical oxygen demand test (sometimes referred to as the **biological oxygen demand** test, or BOD_5). This is a measure of the oxygen required by bacteria and other micro-organisms to break down organic matter in a water sample. It is an indirect measure of the amount of organic matter in a water sample, and gives an indication of how much dissolved oxygen could be removed from water as the organic matter decays.

The test is simple to perform and easily replicable. A sample of water needs to be taken, placed in a clean, darkened glass bottle and left to reach 20°C. Once this has occurred the dissolved oxygen content should be measured (as a concentration). The sample should then be left at 20°C for five days in a darkened environment. After this the dissolved oxygen content should be measured again. The difference between the two dissolved oxygen readings is the BOD_5 value. Over an extended period the dissolved oxygen content of a polluted water sample will look something like that shown in Figure 7.4. In this case the dissolved oxygen content has dropped from 9.0 on day one to 3.6 on day five, giving a BOD_5 value of 5.4 mg/l. After a long period of time (normally more than five days) oxygen will start to be consumed by nitrifying bacteria. In this case the bacteria will be consuming oxygen to turn nitrogenous compounds (e.g. ammonium ions) into nitrate. In order to be sure that nitrifying bacteria are not adding to the oxygen demand a suppressant (commonly allyl thiourea or ATU) is added. This ensures that all the oxygen demand is from the decomposition of organic matter. The use of a five-day period is another safeguard, as, due to the slow growth of nitrifying bacteria, their effect is not noticeable until between eight and ten days (Tebbutt, 1993). There is an argument to be made saying that it does not matter which bacteria are causing the oxygen demand, the test should be looking at all oxygen demand over a five-day period and therefore there is no need to add ATU. However the standard BOD test uses ATU to suppress the nitrifying bacteria.

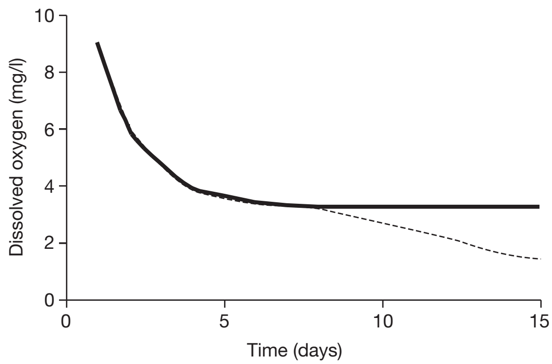

Figure 7.4 Dissolved oxygen curve. The solid line indicates the dissolved oxygen content decreasing due to organic matter. The broken line shows the effect of nitrifying bacteria.

In some cases, particularly when dealing with waste water, the oxygen demand will be higher than total saturation. In this case the sample needs to be diluted with distilled water. The maximum dissolved oxygen content at 20°C is 9.1 mg/l, so any water sample with a BOD_5 value higher than 9 will require dilution. After the diluted test a calculation needs to be performed to find the actual oxygen demand. If you have diluted the sample by half then you need to double your measured BOD_5 value, and so on.

A normal unpolluted stream should have a BOD_5 value of less than 5 mg/l. Untreated sewage is somewhere between 220 and 500 mg/l; while milk has a BOD_5 value of 140,000 mg/l. From these values it is possible to see why a spillage of milk into a stream can have such detrimental effects on the aquatic fauna. The milk is not toxic in its own right, but bacteria consuming the milk will strip the water of any dissolved oxygen and therefore deprive fish of the opportunity to breathe.

There are three reasons why BOD_5 is such a crucial test for water quality:

- Dissolved oxygen is critical to aquatic fauna and the ability to lose dissolved oxygen through organic matter decay is an important measure of stream health.
- It is an indirect measure of the amount of organic matter in the water sample.
- It is the most frequently measured water quality test and has become a standard measure; this means that there are plenty of data to compare readings against.

It also important to realise that BOD is not a direct measure of pollution; rather, it measures the effects of pollution. It also should be borne in mind that there may be other substances present in your water sample that inhibit the natural bacteria (e.g. toxins). In this case the BOD_5 reading may be low despite a high organic load.

Trace organics

Over six hundred organic compounds have been detected in river water, mostly from human activity (Tebbutt, 1993). Examples include benzene, chlorophenols, pesticides, trihalomethanes and polynuclear aromatic hydrocarbons (PAH). These would normally be found in extremely low concentrations but do present significant health risks over the long term. The data for pesticide concentrations (see Table 7.4) in European water resources show that it is a significant problem. This indicates that all water extracted from surface water supplies in Belgium (supplies approximately 30 per cent of the Belgian population) will require pesticide removal before reticulation to customers (Eureau, 2001). Although Germany appears to have no pesticide problem, 10 per cent of its surface water resources occasionally have pesticide levels greater than 0.1 μg/l and 90 per cent have pesticides in concentrations less than 0.1 μg/l (but still present) (Eureau, 2001).

Table 7.4 Percentage of water resources with pesticide concentrations regularly greater than 0.1 μg/l (European Union drinking water standard) for selected European countries

Country	*Surface water (%)*	*Groundwater (%)*
Belgium	100	5.2
Denmark	n/a	8.9
Germany	0.0	0.0
Netherlands	50.0	5.0
UK	77.0	6.0

Source: Data from Eureau (2001)

Some of the trace organic compounds accumulate through the food chain so that humans and other species that eat large aquatic fauna may be at risk. Of particular concern are endocrine disrupting chemicals (EDCs), which have been detected in many rivers. These chemicals, mostly a by-product of industrial processes, attack the endocrine system of humans and other mammals, affecting hormone levels. Some chemicals (e.g. DDT) have the ability to mimic the natural hormone oestrogen. Because oestrogen is part of the reproductive process these chemicals have the potential to affect reproductive organs and even DNA. Studies have shown high levels of oestrogen-mimicking compounds in sewage effluent (Montagnani *et al*., 1996) and that male fish held in cages at sewage effluent discharge sites can develop female sexual organs (Jobling and Sumpter, 1993).

Trace organics can be detected using gas chromatography, although this is made difficult by the sheer number of compounds to be detected. They are removed from drinking water supplies using activated carbon filters, or sometimes oxidation by ozone.

Nitrogen compounds

Nitrogen exists in the freshwater environment in four main forms:

- organic nitrogen – proteins, amino acids and urea
- ammonia – either as free ammonia (NH_3) or the ammonium ion (NH_4^+)
- nitrite (NO_2^-)
- nitrate (NO_3^{2-}).

If organic nitrogen compounds enter a river (e.g. in untreated sewage) then an oxidation process called nitrification takes places. An approximation of the process is outlined below:

$$\text{Organic N} + O_2 \rightarrow NH_3/NH_4^+ + O_2 \rightarrow NO_2^- + O_2 \rightarrow NO_3^{2-}$$

For this to occur there must be nitrifying bacteria and oxygen present. This is one of the main processes operating in a sewage treatment works (see pp. 143–145) – the breakdown of organic nitrogenous compounds into a stable and relatively harmless nitrate. There are two problems with this process occurring in the natural river environment. First, there is the oxygen demand created by the nitrification process. Second, the intermediate ammonia stage is highly toxic, even in very low concentrations. Under extremely low dissolved oxygen concentrations (less than 1 mg/l) the nitrification process can be reversed, at least in the first stage. In this case nitrates will turn into nitrite and oxygen will be released. Unfortunately, this is not a ready means for re-oxygenating a river as by the time the dissolved oxygen level has dropped to 1 mg/l the fish population will have died or moved elsewhere.

The levels of nitrate in a water sample can be expressed in two different ways: absolute nitrate concentration, or the amount of nitrogen held as nitrate (normally denoted as NO_3–N). The two are related by a constant value of approximately 4.4. As an example the World Health Organisation recommended that the drinking water standard for nitrate in drinking water be 45 mg/l. This can also be expressed as 10 mg/l NO_3–N.

As indicated above, one source of nitrate is from treated sewage. A second source is from agricultural fertilisers. Farmers apply nitrate fertilisers to enhance plant growth, particularly during the spring. Plants require nitrogen to produce green leaves, and nitrates are the easiest form to apply as a fertiliser. This is because nitrates are extremely soluble and can easily be taken up by the plant through its root system. Unfortunately this high solubility makes them liable to be flushed through the soil water system and into rivers. To make matters worse a popular fertiliser is ammonium nitrate – $(NH_4)_2NO_3$. This has the added advantage for the farmer of three nitrogen atoms per molecule. It has the disadvantage for the freshwater environment of extremely high solubility and providing ammonium ions in addition to nitrate. The application of nitrate fertilisers is most common in areas of intensive agricultural production such as arable and intensive livestock farming.

Another source of nitrates in river systems is from animal wastes, particularly in dairy farming where slurry is applied to fields. This is organic nitrogen (frequently with high urea content from urine) which will break down to form nitrates. This is part of the nitrification process described earlier.

A fourth source of nitrates in river systems is from plants that capture nitrogen gas from the air. This is not strictly true, as it is actually bacteria such as Rhizobium, attached to a plant's root, that capture the gaseous nitrogen and turn it into water-soluble forms for the plants to use. Not all plants have this ability; in agriculture it is the legumes, such as clovers, lucerne (or alfalfa), peas and soy beans, that can gain nitrogen in this way. Once the nitrogen is in a soluble form it can leach through to a river system in the same way that fertilisers do. Over a summer period the nitrogen levels in a soil build up and then are washed out when autumn and winter rains arrive. This effect is exacerbated by ploughing in the autumn, which releases large amounts of soil-bound nitrogen.

There is one other source of nitrates in rivers: atmospheric pollution. Nitrogen gas (the largest constituent of the atmosphere) will combine with oxygen whenever there is enough energy for it to do so. This energy is readily supplied by combustion engines (cars, trucks, industry, etc.) producing various forms of nitrogen oxide gases (often referred to as NO_x gases). These gases are soluble to water in the atmosphere and form nitrites and nitrates in rainwater. This is not a well-studied area and it is difficult to quantify how much nitrogen reaches rivers from this source (see p. 35).

The different sources of nitrate in a river add together to give a cycle of levels to be expected in a year. Figure 7.5 shows this cycle over a three-year period on the river Lea, south-east England. The low points of nitrate levels correspond to the end of a summer period, with distinct peaks being visible over the autumn to spring period, particularly in the spring. The Lea is a river that has intensive arable agriculture in its upper reaches, but also a significant input from sewage effluent. At times during the summer months the Lea can consist of completely recycled water, and the water may have been through more than one sewage works. This gives a background nitrate level, but it is perhaps surprising that the summer levels of nitrate are not higher, compared to the winter period. Partly this can be attributed to the growth of aquatic plants in the summer, which remove nitrate from the water. The peaks over the autumn–spring period are as a result of agricultural practices discussed above. The example given here is specific to the south-east of England; in different parts of the world the cycles will differ in timing and extent.

Nitrates are relatively inert and do not create a major health concern. An exception to this is methaemoglobinaemina ('blue baby syndrome'). Newborn babies do not have the bacteria in their stomach to deal with nitrates in the same manner as older children and adults. In the reducing surroundings of the stomach the nitrate is transformed into nitrite that then attaches itself to the haemoglobin molecule in red blood cells, preferentially replacing oxygen. This leads to a reduction in oxygen supply around the body, hence the name 'blue baby syndrome'. In reality methaemoglobinaemina is extremely rare, possibly coming from nitrate-polluted well supplies but not mains-supplied drinking water. The drinking water limit for the European Union is 50 mg/l of nitrate (44 mg/l in the USA). In rivers it is rare to have nitrate values as high as this. In a study of streams draining intensively dairy-farmed land in the North Island of New Zealand, Rodda *et al*. (1999) report maximum nitrate levels of 26.4 mg/l. These are reported as being 'very high by New Zealand standards' (Rodda *et al*., 1999: 77). In Figure 7.5 the peak nitrate level for the river Lea in England is 21 mg/l, with the norm being somewhere between 5 and 10 mg/l.

The biggest concern with nitrates in a river system is **eutrophication**. In exactly the same way that the nitrogen enhances the growth of land-based plants, it will also boost the growth of aquatic plants, including algae. This creates a problem of over-production of plant matter in river systems. This is discussed in more detail on pp. 142–143.

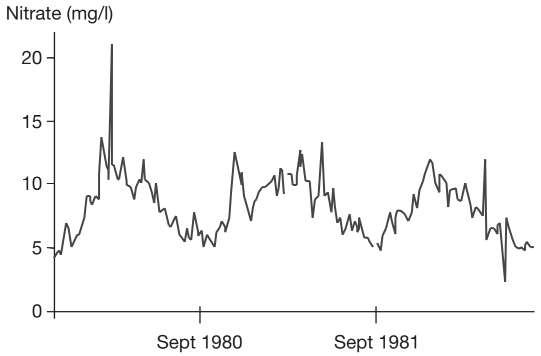

Figure 7.5 Nitrate levels in the river Lea, England. Three years of records are shown: from September 1979 until September 1982.

Source: Data from the Environment Agency

Phosphates

Phosphorus can be found in three different forms: orthophosphate, polyphosphate (both normally dissolved) and organic phosphate (bound to organic particles). The ratio of different forms of phosphorus in a water sample is highly pH dependent (Chapman, 1996). Like nitrogen, the availability of phosphorus is a limiting factor in plant growth. The most common form of application for plants is as phosphate. The major difference from nitrates is that phosphate is not nearly as soluble. Consequently phosphate is normally applied as a solid fertiliser, and less frequently than nitrate. In river systems the main source of dissolved phosphate is from detergents and soaps that come through sewage treatment works. Sewage treatment works remove very little of the phosphate from detergents present in waste water, except where specific phosphate-stripping units are used. The largest amount of phosphate in river systems is normally attached to particles of sediment. Rodda *et al*. (1999) report maximum dissolved reactive phosphorus levels of 0.2 mg/l but total phosphorus levels of 1.6 mg/l. This is for intensive dairy production, where the majority of phosphate is from agricultural fertilisers.

Phosphates are a major contributor to eutrophication problems. The fact that they are bound to sediments means that they often stay in a river system for a long period of time. Improvements in water quality for a river can often be delayed substantially by the steady release of phosphate from sediments on the river bed.

Chlorine

Chlorine is not normally found in river water. It is used as a disinfectant in the supply of drinking water. It is used because it is toxic to bacteria and relatively short lived. More common to find in river water samples is the chloride ion. This may be an indicator of sewage pollution as there is a high chloride content in urine. Chloride ions give the brackish taste of sea water, the threshold for taste being around 300 mg/l. The European Commission limit for drinking water is 200 mg/l.

Heavy metals

'Heavy metals' is the term applied to metals with an atomic weight greater than 6. They are generally only found in very low levels dissolved in fresh water, but may be found in bed load sediments. In acidic waters metals can be dissolved (i.e. found in ionic form). They are often toxic in concentrations above trace levels. The toxicity, in decreasing order, is mercury, cadmium, copper, zinc, nickel, lead, chromium, aluminium and cobalt (Gray, 1999). In the aquatic environment copper and zinc are the most frequent causes of toxicity. A major source of zinc is derived from galvanised steel, particularly in wire fencing and roofs (Alloway and Ayres, 1997).

Accumulation of lead in sediments has been a problem for aquatic wildlife. Since the banning of leaded petrol the major source has been through the use of lead shot and fishing sinkers. Lead shot has been banned in favour of steel shot in many countries (e.g. USA, UK, New Zealand, Australia) due to these problems (Dodds, 2002).

The sources of heavy metals in the aquatic environment are almost always industrial or surface runoff from roads. Sewage sludge (the product of sedimentation at a sewage treatment works) is frequently heavy metal-rich, derived from industry discharging waste into the sewerage system. When untreated sewage is discharged into a river heavy metals can be found in the sediments. Where there is a combined sewage and storm-water drainage system for an urban area, untreated sewage can be discharged during a storm event when the sewage treatment works cannot cope with the extra storm water. Runoff from roads (through a stormwater system) frequently shows high levels of copper from vehicle brake pads. When washed into a river system, particularly in summer storms, the copper levels can be extremely high and cause toxicity problems to aquatic fauna.

WATER-QUALITY MEASUREMENT

The techniques used for water-quality analysis vary considerably depending on equipment available and the accuracy of measurement required. For the highest accuracy of measurement water samples should be taken back to a laboratory, but this is not always feasible. There are methods that can be carried out in the field to gain a rapid assessment of water quality. Both field and laboratory techniques are discussed on the following pages. Before discussing the measurement techniques it is important to consider how to sample for water quality.

Sampling methodology

It is difficult to be specific on how frequently a water sample should be taken, or how many samples represent a given stretch of water. The best way of finding this out is to take as many measurements as possible in a trial run. Then statistical analyses can be carried out to see how much difference it would have made to have had fewer measurements. By working backwards from a large data set it is possible to deduce how few measurements can be taken while still maintaining some accuracy of overall assessment. An example of this type of approach, when used for the reduction in a hydrometry (water quantity) network, is in Pearson (1998). The main concern is that there are enough measurements to capture the temporal variability present and that the sample site is adequately representative of your river stretch.

One important consideration that needs to be understood is that the sample of water taken at a particular site is representative of all the catchment above it, not just the land use immediately adjacent. Adjacent land use may have some influence on the water quality of a sample, but this will be in addition to any affect from land uses further upstream which may be more significant.

Gravimetric methods

Gravimetric analysis depends on the weighing of solids obtained from a sample by evaporation, filtration or precipitation (or a combination of these three). This requires an extremely accurate weighing balance and a drying oven, hence it is a laboratory technique rather than a field one. An example of gravimetric analysis is the standard method for measuring total dissolved solids (TDS). This is to filter a known volume of water through 0.45 μm (1 micron = one-millionth of a metre or one-thousandth of a millimetre) filter paper. The sample of water is then dried at 105°C and the weight of residue left is the TDS.

Other examples of gravimetric analysis are total suspended solids and sulphates (causing a precipitate and then weighing it).

Volumetric methods

Volumetric analysis is using titration techniques to find concentrations of designated substances. It is dependent on measuring the volume of a liquid reagent (of known concentration) that causes a visible chemical reaction. This is another laboratory technique as it requires accurate measurements of volume using pipettes and burettes. Examples of this technique are chloride and dissolved oxygen (using the Winkler method).

Colorimetry

Colorimetric analysis depends on a reagent causing a colour to be formed when reacting with the particular ion you are interested in measuring. The strength of colour produced is assumed to be proportional to the concentration of the ion being measured (Beer's law). The strength of colour can then be assessed using one of four techniques: comparison tubes, colour discs, colorimeter or spectrophotometer.

Comparison tubes are prepared by using standard solutions of the ion under investigation which the reagent is added to. By having a range of standard solutions the strength of colour can be compared (by eye) to find the concentration of the water sample. The standard solutions will fade with time and need remaking, hence this is a time-consuming method.

Colour discs use the same principle as comparison tubes, except in this case the standards are in the form of coloured glass or plastic filters. The coloured sample is visually compared to the coloured disc to find the corresponding concentration. It is possible to buy colour disc kits that come with small packets of reagent powder for assessment of a particular ion. This method is extremely convenient for rapid field assessment, but is subjective and prone to inaccuracy.

A colorimeter (sometimes called an absorptiometer) takes the subjective element out of the assessment. It is similar to a turbidity meter in that a beam of light is shone through the reagent in a test tube. The amount of light emerging from the other side is detected by a photo-electric cell. The darker the solution (caused by a high concentration of reactive ion) the less light emerges. This reading can then be compared against calibrations done for standard solutions.

A spectrophotometer is the most sophisticated form of colorimetric assessment. In this case instead of a beam of white light being shone through the sample (as for the colorimeter) a specific wavelength of light is chosen. The wavelength chosen will depend on the colour generated by the reagent and is specified by the reagent's manufacturer.

There are a range of spectrophotometers available to perform rapid analysis of water quality in either a laboratory or field situation. Many ions of interest in water-quality analysis can be assessed using colorimetric analysis. These include nitrate, nitrite, ammonia and phosphate.

Ion-selective electrodes

In a similar vein to pH meters ion-selective electrodes detect particular ions in solution and measure the electrical potential produced between two reactive substances. The tip of the electrode in the instrument has to be coated with a substance that reacts with the selected ion. With time the reactive ability of the electrode will decrease and need to be replaced. Although convenient for field usage and accurate, the constant need for replacing electrodes makes these an expensive item to maintain. There are ion-selective electrodes available to measure dissolved oxygen, ammonium, nitrate, calcium, chloride and others.

Spectral techniques

When ions are energised by passing electricity through them, or in a flame, they produce distinctive colours. For instance, sodium produces a distinctive yellow colour, as evidenced by sodium lamps used in some cars and street lamps. Using spectral analysis techniques the light intensity of particular ions in a flame are measured and compared to the light intensity from known standard solutions. The most common form of this analysis is atomic absorption spectrophotometry, a laboratory technique which is mostly used for metallic ions.

PROXY MEASURES OF WATER QUALITY

Any measurement of water quality using individual parameters is vulnerable to the accusation that it represents one particular point of time but not the overall water quality. It is often more sensible to try and assess water quality through indirect measurement of something else that we know is influenced by water quality. Two such proxy measures of water quality are provided by biological indicators and analysis of sediments in the river.

Biological indicators

Aquatic fauna normally remain within a stretch of water and have to try and tolerate whatever water pollution may be present. Consequently the health of aquatic fauna gives a very good indication of the water quality through a reasonable period of time. There are two different ways that this can be done: catching fauna and assessing their health; or looking for the presence and absence of key indicator species.

Fish surveys are a common method used for assessing the overall water quality in a river. It is an expensive field technique as it requires substantial human resources: people to wade through the water with electric stun guns and then weigh and measure stunned fish. When this is done regularly it gives very good background information on the overall water quality of a river.

More common are biological surveys using indicator species, particularly of macro-invertebrates. Kick sampling uses this technique. A bottom-based net is kicked into sediment to catch any bottom-based macro-invertebrates, which are then counted and identified. There are numerous methods that can be used to collate this species information. In Britain the BMWP (Biological Monitoring Working Party) score is commonly used and provides good results. Species are given a score ranging from 1 to 10, with 10 representing species that are extremely intolerant to pollution. The presence of any species is scored (it is purely presence/absence, not the total number) and the total for the kick sample calculated. The BMWP score has a maximum of 250. Other indicator species scores include the Chandler index and the ASPT (Average Score Per Taxon). Details of these can be found in a more detailed water-quality assessment text such as Chapman (1996).

Another example of an indicator species used for water-quality testing is *Escherichia coli* (*E. coli*). These are used to indicate the presence or absence of faecal contamination in water. *E. coli* is a bacteria present in the intestines of all mammals and excreted in large numbers in faeces. Although one particular strain (*E. coli*$_{157}$) has toxic side effects the vast majority of *E. coli* are harmless to humans. Their presence in a water sample is indicative of faecal pollution, which may be dangerous because of other pathogens carried in the contaminated water. They are used as an indicator species because they are easy to detect, while viruses and other pathogens are extremely difficult to measure. Coliform bacteria (i.e. bacteria of the intestine) are detected by their ability to ferment lactose, producing acid and gas (Tebbutt, 1993). There are specific tests to grow *E. coli* in a lactose medium, which allow the tester to derive the most probable number per 100 ml (MPN/100 ml).

Sediments

The water in a channel is not the only part of a river that may be affected by water pollution. There are many substances that can build up in the sediments at the bottom of a river and provide a record of pollution. There are two big advantages to this method for investigating water quality: the sediments will reflect both instantaneous large pollution events and long, slow contamination at low levels; and if the river is particularly calm in a certain location the sediment provides a record of pollution with time (i.e. depth equals time). Not all water pollutants will stay in sediments, but some are particularly well suited to study in this manner (e.g. heavy metals and phosphorus).

The interpretation of results is made difficult by the mobility of some pollutants within sediments. Some metals will bind very strongly to clay particles in the sediments (e.g. lead and copper), and you can be fairly certain that their position is indicative of where they were deposited. Others will readily disassociate from the particles and move around in the interstitial water (e.g. zinc and cadmium)

(Alloway and Ayres, 1997). In this case you cannot be sure that a particularly high reading at one depth is from deposition at any particular time.

MODELLING WATER QUALITY

The numerical modelling of water quality is frequently required, particularly to investigate the effects of particular water-quality scenarios. The type of problems investigated by modelling are: the impact of certain levels of waste discharge on a river (particularly under low flow levels); recovery of a water body after a pollution event; the role of backwaters for concentration of pollutants in a river; and many more. The simplest water-quality models look at the concentration of a certain pollutant in a river given knowledge about flow conditions and decay rates of the pollutant. The degradation of a pollutant with time can be simulated as a simple exponential decay rate equation. A simple mass balance approach can then be used to calculate the amount of pollutant left in the river after a given period of time (James, 1993). More complex models build on this approach and incorporate ideas of diffusion, critical loads of pollutants and chemical reaction between pollutants in a river system. If the problem being researched is to track pollutants down a river then it is necessary to incorporate two- or three-dimensional representation of flow hydraulics. There are numerous water-quality models available in the research literature, as well as those used by consultants and water managers.

EUTROPHICATION

'Eutrophication' is the term used to describe the addition of nutrients to an aquatic ecosystem that leads to an increase in net primary productivity. The term comes from limnology (the study of freshwater bodies, e.g. lakes and ponds) and is part of an overall classification system for the nutrition, or trophic, level of a freshwater body. The general classification moves from oligotrophic (literally 'few nutrients'), to eutrophic ('good nutrition') and ends with hypertrophic ('excess nutrients'). In limnology this classification is viewed as part of a natural progression for bodies of water as they fill up with sediment and plant matter. Eutrophication is a natural process (as part of the nitrogen and phosphorus cycles), but it is the addition of extra nutrients from anthropogenic activity that attracts the main concern in hydrology. In order to distinguish between natural and human-induced processes the term 'cultural eutrophication' is sometimes used to identify the latter.

The major nutrients that restrict the extent of a plant's growth are potassium (K), nitrogen (N) and phosphorus (P). If you buy common fertiliser for a garden you will normally see the K:N:P ratio expressed to indicate the strength of the fertiliser. For both aquatic and terrestrial plants nitrogen is required for the production of chlorophyll and green leaves, while potassium and phosphorus are needed for root and stem growth. In the presence of abundant nitrogen and phosphorus (common water pollutants, see pp. 136–138), aquatic plant growth, including algae, will increase dramatically. This can be seen as positive as it is one way of removing the nitrate and phosphate from the water, but overall it has a negative impact on the river system. The main negative effect is a depletion of dissolved oxygen caused by bacteria decomposing dead vegetative matter in the river. In temperate regions this is a particular problem in the autumn when the aquatic vegetation naturally dies back. In tropical regions it is a continual problem. A second negative effect is from algal blooms. In 1989 there was an explosion in cyano-bacteria numbers in Rutland Water, a reservoir supplying drinking water in central England (Howard, 1994). (NB These are also called blue-green algae, despite being a species of cyano-bacteria.) The cyano-bacteria produce toxins as waste products of respiration that can severely affect water quality. In the 1989 outbreak several dogs and sheep that drank water from Rutland Water were poisoned, although no humans were affected (Howard, 1994). In an effort to eliminate future problems the nutrient-rich source water for Rutland

Water is supplemented with water from purer river water pumped from further afield.

Eutrophication of water can occur at what appear to be very low nutrient levels. As an example the drinking water standard for nitrate-nitrogen is around 12 mg/l (depending on country) but concentrations as low as 2–3 mg/l can cause eutrophication problems in water bodies.

Table 7.5 shows some of the indicators used in a quantitative example of defined trophic levels developed for the Organisation for Economic Co-operation and Development (OECD). The chlorophyll is an indicator of algal growth in the water, while phosphorus and dissolved oxygen are more traditional water-quality measures. The dissolved oxygen is taken from the bottom of the lake because this is where the vegetative decomposition is taking place. The dissolved oxygen level near to the surface will vary more because of the proximity to the water/air interface and the oxygen produced in photosynthesis by aquatic plants. It is worth noting that heavily eutrophied water samples will sometimes have a dissolved oxygen greater than 100 per cent. This is due to the oxygen being produced by algae which can supersaturate the water.

CONTROLLING WATER QUALITY

Waste water treatment

The treatment of waste water is a relatively simple process that mimics natural processes in a controlled, unnatural environment. The treatment processes used for industrial waste water is dependent on the type of waste being produced. In this section the processes described are those generally found in sewage treatment rather than in specialised industrial waste water treatment.

There are two major objectives for successful sewage treatment: to control the spread of disease from waste products and to break down the organic waste products into relatively harmless metabolites (i.e. by-products of metabolism by bacteria, etc.). The first objective is achieved by isolating the waste away from animal hosts so that viruses and other pathogens die. The second objective is particularly important for the protection of where the treated effluent ends up – frequently a river environment.

In Britain the first attempt to give guidelines for standards of sewage effluent discharge were provided by the Royal Commission on Sewage Disposal which sat between 1898 and 1915. The guidelines are based on two water-quality parameters described earlier in this chapter: suspended solids and biochemical oxygen demand (BOD). The Royal Commission set the so-called 30:20 standard which is still applicable today (i.e. 30 mg/l of suspended solids and 20 mg/l of BOD). The standard was based on a dilution ratio of 8:1 with river water. Where river flow is greater than eight times the amount of sewage effluent discharge the effluent should have a TSS of less than 30 mg/l and a BOD of less than 20 mg/l. There was also the recommendation that if the river is used for drinking water extraction further downstream the standard should be tightened to 10:10. This was

Table 7.5 OECD classification of lakes and reservoirs for temperate climates

Trophic level	*Average total P (mg/l)*	*Dissolved oxygen (% saturation)*	*Max. chlorophyll (mg/l) (at depth)*
Ultra-oligotrophic	0.004	>90	0.0025
Oligotrophic	0.01	>80	0.008
Mesotrophic	0.01–0.035	40–89	0.008–0.025
Eutrophic	0.035–0.1	0–40	0.025–0.075
Hypertrophic	>0.1	0–10	>0.075

Source: Adapted from Meybeck *et al.* (1989)

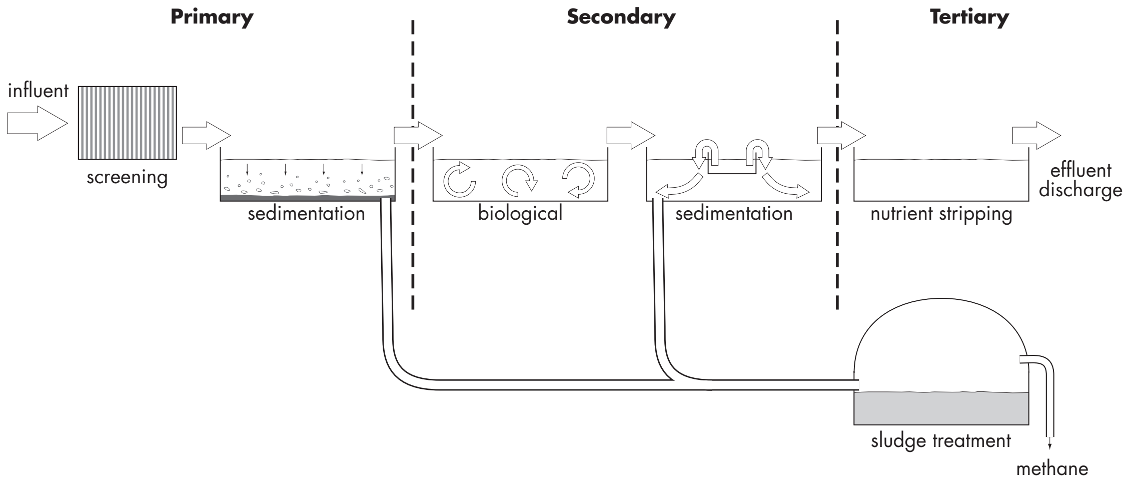

Figure 7.6 Schematic representation of waste water treatment from primary through to tertiary treatment, and discharge of the liquid effluent into a river, lake or the sea.

used as a recommendation until the 1970s when a system of legal consents to discharge was introduced (see p. 147).

The processes operating at a waste water treatment works are very simple. They are summarised below and in Figure 7.6 (NB not every sewage treatment works will have all of these processes present).

1 Primary treatment: screening and initial settlement.
2 Secondary treatment: encouraging the biological breakdown of waste and settling out of remaining solids. This can take place either in trickle bed filters or activated sludge tanks. The main requirement is plenty of oxygen to allow micro-organisms to break down the concentrated effluent.
3 Tertiary treatment: biodigestion of sludge (from earlier settling treatment); extra treatment of discharging effluent to meet water-quality standards (e.g. phosphate stripping, nitrate reduction).

Raw sewage entering a sewage treatment works is approximately 99.9 per cent water (Gray, 1999). This is derived from water used in washing and toilet flushing, and also from storm runoff in an urban environment where there is a combined sewage/stormwater drainage scheme. Of the solids involved, the majority are organic and about half are dissolved in the water (TDS). Of the organic compounds the breakdown is approximately 65 per cent nitrogenous (proteins and urea), 25 per cent carbohydrates (sugars, starches, cellulose) and 10 per cent fats (cooking oils, grease, soaps) (Gray, 1999). Typical values for TSS and BOD at different stages of sewage treatment are provided in Table 7.6.

In tertiary treatment an effort is sometimes (but not always) made to reduce the level of nitrate and phosphorus in the discharged waste. In some cases this is achieved through final settling ponds where the growth of aquatic flora is encouraged and the nutrients are taken up by the plants before discharge into a stream. Of particular use are reeds which do not die back during the winter (in temperate regions). This is a re-creation of natural wetlands that have been shown to be extremely efficient removers of both nitrogen and phosphorus from streams (e.g. Russel and Maltby, 1995). Other methods of phosphate removal are to add a lime or metallic salt coagulant that causes a chemical

Table 7.6 Changes in suspended solids and biochemical oxygen demand through sewage treatment. These are typical values which will vary considerably between treatment works

Stage of treatment	*Suspended solids (mg/l)*	*BOD (mg/l)*
Raw sewage	400	300
After primary treatment	150	200
After biological treatment	300	20
Effluent discharged to river	30	20

reaction with the dissolved phosphorus so that an insoluble form of phosphate settles out. This is particularly useful where the receiving water for the final effluent has problems with eutrophication. The average phosphorus concentration in raw sewage is 5–20 mg/l, of which only 1–2 mg/l is removed in biological treatment.

In some cases, particularly in the USA, chlorination of the discharging effluent can take place. Chlorine is used as a disinfectant to kill any pathogens left after sewage treatment. This is a noble aim but creates its own difficulties. The chlorine can attach to organic matter left in the effluent and create far worse substances such as polychlorinated biphenyl (PCB) compounds. Another safer form of disinfection is to use ultraviolet light, although this can be expensive to install and maintain.

Source control

The best way of controlling any pollution is to try and prevent it happening in the first place. In order to achieve this differentiation has to be made between point source and diffuse pollutants (see p. 129). When control over the source of pollutants is achieved dramatic improvements in river-water quality can be achieved. An example of this is shown in the Case Study of the Nashua River in Massachusetts, USA.

Case study

CONTROLLING WATER QUALITY OF THE NASHUA RIVER

The Nashua river is an aquatic ecosystem that has undergone remarkable change in the last one hundred years. It drains an area of approximately 1,400 km^2 in the state of Massachusetts, USA, and is a tributary of the much larger Merrimack river which eventually flows into the sea in Boston Harbor (see Figure 7.7). The land use of the Nashua catchment is predominantly forest and agricultural, with a series of towns along the river. It is the industry associated with these towns that has brought about the changes in the Nashua, predominantly through the twentieth century. The latter-day changes are well illustrated by the two photographs at the same stretch of the Nashua, in 1965 and 1995 (see Plates 9 and 10).

Prior to European colonisation of North America the Nashua valley was home to the Nashaway tribe, and the Nashua river could be considered to be in a pristine condition. With the arrival of European settlers to New England the area was used for agriculture and the saw milling of the extensive forests. The Industrial Revolution of the nineteenth century brought manufacturing to the area and mills sprang up along the river. By the middle of the twentieth century the small towns along the Nashua (Gardner, Fitchburg, Leominster and Nashua) were home to paper, textile and shoe factories, many of which were extracting water from the river and then discharging untreated waste back into the river. The photograph of the Nashua in 1965 (Plate 9) is indicative of the pollution problems experienced in the river; in this case dye from a local paper factory has turned the river red. Under the US water-quality classification scheme the river was classified as U: unfit to receive further sewage.

In 1965 the Nashua River Clean-Up Committee was set up to try to instigate a plan of restoring the water quality in the river. This committee later became the Nashua River Watershed Association (NRWA) which still works today to improve water-quality standards in the area. Between 1972 and 1991 eleven waste water-treatment plants were constructed or upgraded to treat waste from domestic, and to a lesser extent from industrial, sources in the catchment. These were built using grants from the state and federal government as part of a strategy to improve the river from U to B status (fit for fishing and swimming). Through this control of point source pollution the river-water quality has improved dramatically as can be seen in the second photograph of the river (Plate 10). The river has attained B status and is an important recreational asset for the region. It has not returned to a pristine state, though, and is unlikely to while there is still a significant urban population in the catchment. There are problems with combined sewage and stormwater drainage systems discharging untreated waste into the river during large storms, and also diffuse pollution sources – particularly in the urban environment. However, during the latter half of the twentieth century the Nashua river has had its water quality transformed from an abiotic sewer into a clean river capable of maintaining a healthy salmonid fish population. This has largely been achieved through the control of point pollution sources.

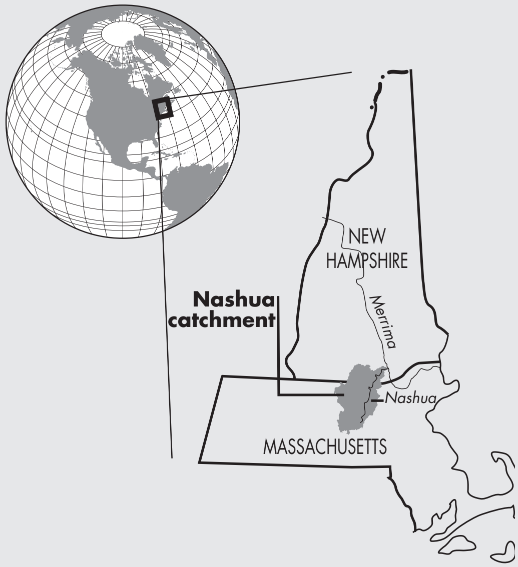

Figure 7.7 Location of the Nashua catchment in north-east USA.

The author gratefully acknowledges the Nashua River Watershed Association for supplying much of this information and Plates 9 and 10. For more information on the NRWA visit: http://www.nashuariverwatershed.org/

Controlling point source pollutants

The control of point source pollutants cannot always be achieved by removing that point source. It is part of water resource management to recognise that there may be valid reasons for disposing of waste in a river; effective management ensures that waste disposal creates no harmful side effects. In the United Kingdom the control of point source pollution is through discharge consents. These provide a legal limit for worst-case scenarios – for example, at individual sewage treatment works they are usually set with respect to TSS, BOD and ammonia (sometimes heavy metals are included), and calculated to allow

Technique: Calculating discharge consents

In England and Wales the setting of discharge consents for point source pollution control is carried out by the Environment Agency. A discharge consent gives a company the right to dispose of a certain amount of liquid waste into a river system so long as the pollution levels within the discharge are below certain levels. To calculate what those critical levels are a series of computer programs are used. These computer programs are in the public domain and can be obtained from the Environment Agency. They use very simple principles that are described here.

The main part of the discharge consent calculation concentrates on a simple mass balance equation (7.3):

$$C_D = \frac{Q_U C_U + Q_E C_E}{Q_B + Q_E} \tag{7.3}$$

where Q refers to the amount of flow (m^3/s) and C the concentration of pollutant. For the subscripts: D is for downstream; U is for upstream (i.e. the background); and E is for the effluent.

With this mass balance equation the downstream concentration can be calculated with varying flows and levels of effluent concentrations. This variation in flow and concentration is achieved through a computer program running a Monte Carlo simulation.

In this case the Monte Carlo simulation involves a random series of values for Q_U, Q_E, C_U and C_E drawn from an assumed distribution for each variable. It is assumed that the distributions are log-normal in shape (see Figure 7.8) and therefore using the data in Table 7.7 the actual distribution for each variable is simulated. Once the distribution for each variable is known then a random variable is chosen from that distribution. In the case of a log-normal distribution this means that it is most likely to be close to the mean value but more likely to be above than below the mean (see Figure 7.8). In a Monte Carlo simulation the value of C_D is calculated many times (often set to 1,000) so that a distribution for C_D can be drawn. The consent to discharge figure is taken from the distribution of C_D, usually looking at the 90 or 95-percentile values, i.e. the target will be achieved 90 or 95 per cent of the time.

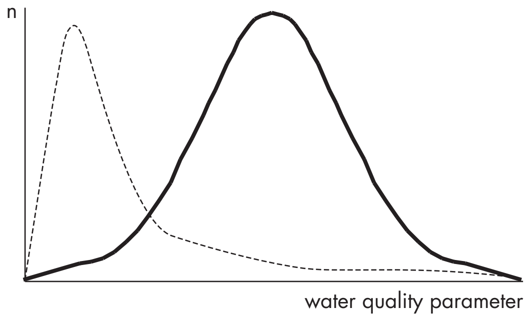

Figure 7.8 A log-normal distribution (broken line) compared to a normal distribution (solid line).

Table 7.7 Parameters required to run a Monte Carlo simulation to assess a discharge consent

Variable	*Required data*
River flow (Q_U)	Mean daily flow and Q_{95}
Upstream river quality (C_U)	Mean value and standard deviation
Effluent flow (Q_E)	Mean value and standard deviation
Effluent quality (C_E)	Mean and standard deviation

The values that are required for calculating a consent to discharge (see Table 7.7) are derived from normal hydrological data. River flow data can be derived from a flow duration curve (see Chapter 6). The water quality information requires at least three to four years of regular measurements. The values to describe Q_E and C_E will either be known or are to be varied in the simulation in order to derive a consent to discharge value.

In short, the person calculating the discharge consent inserts values from Table 7.7 into the Monte Carlo simulation. This will then produce the 95 percentile value of C_D. If that value is too high (i.e. too much pollution) then the simulation is run again using lower values for C_E until a reasonable value is derived. Once the reasonable value has been reached then the 95 percentile value of C_E is taken as the discharge consent. The definition of a 'reasonable value' will be dependent on the designated use of the river. Rivers with high-class fisheries and those with abstraction for potable supply have much higher standards than for other uses.

The approach described here can be used to calculate a consent to discharge for such water quality parameters as BOD and TSS. When a calculation is being carried out for ammonia then more data are required to describe the water quality in the receiving river. Parameters such as pH, temperature, alkalinity, TDS and dissolved oxygen (all described with mean and standard deviation values) are required so that chemical reaction rates within the river can be calculated.

A scheme such as discharge consents provides a legal framework for the control of point source pollutants but the actual control comes about through implementing improved waste water treatment.

for low flow levels in the receiving stream (see the technique box for calculating discharge consents on p. 147). There is also an obligation to comply with the European Union Urban Waste Water directive.

Controlling diffuse source pollutants

The control of diffuse source water pollution is much harder to achieve. In an urban environment this can be achieved through the collection of stormwater drainage into artificial wetlands where natural processes can lessen the impact of the pollutants on the draining stream. Of particular concern is runoff derived from road surfaces where many pollutants are present as waste products from vehicles. Hamilton and Harrison (1991) suggest that although roads only make up 5–8 per cent of an urban catchment area they can contribute up to 50 per cent of the TSS, 50 per cent of the total hydrocarbons and 75 per cent of the total heavy metals input into a stream. The highest pollutant loading comes during long, dry periods which may be broken by flushes of high rainfall (e.g. summer months in temperate regions). In this case the majority of pollutants reach the stream in the first flush of runoff. If this runoff can be captured and held then the impact of these diffuse pollutants is lessened. This is common practice for motorway runoff where it drains into a holding pond before moving into a nearby water course.

Another management tool for control of diffuse pollutants is to place restrictions on land management practices. An example of this is in areas of England that have been designated either a Nitrate Vulnerable Zone (NVZ) or a Nitrate Sensitive Area (NSA), predominantly through fears of nitrate contamination in aquifers. In NSAs the agricultural practices of muck spreading and fertilising with nitrates are heavily restricted. This type of control relies on tight implementation of land use planning – something that is not found uniformly between countries, or even within countries.

Examples of controlling water pollution

Biggs (1989) presents data showing the recovery of a river in New Zealand following effective treatment of a point source pollution problem. The pollution was due to discharge of untreated effluent from an abattoir, directly into a nearby branch of the Waimakariri river. Water quality was monitored upstream and downstream of the discharge point before and after a staged improvement in wastewater treatment at the abattoir. The results shown in

Figure 7.9 use an autotrophic index, a ratio of the periphyton mass to the chlorophyll-a. This is a measure of the proportions of heterotrophic (require organic carbon to survive) to autotrophic (produce organic compounds from simple molecules) organisms. The time series of data upstream and downstream from the abattoir (Figure 7.9) can be split between the pre-treatment (Sep–Oct 1985), the recovery period (May–Aug 1986) and the recovered period (after August 1986) (Biggs, 1989). (NB the vertical axis is on a logarithmic scale so differences appear smaller.) A remarkable point about this study is how quickly the river appears to have recovered (approximately five months) following treatment of the point source pollution. This is a reflection of the low residence time of the pollutants within the river system and the effective flushing out of the pollutants by the river.

Dodds (2002) presents two case studies on lake eutrophication with varying degrees of success. The first is for Lake Washington on the eastern border of Seattle, USA. For Lake Washington the diversion of treated sewage away from the lake (achieved in 1963) was enough to halt the decline in water quality and return the lake to an oligotrophic state. For Lake Trummen in Sweden the stopping of sewage input into the lake was not enough to improve water quality since high levels of phosphorous remained in the lake sediment continuing the eutrophication problem. In this case a dramatic rise in water quality was achieved by dredging the lake sediments (and selling the dredged sediments as nutrient topsoil) so that the lake was able to be returned to recreational usage (Dodds, 2002).

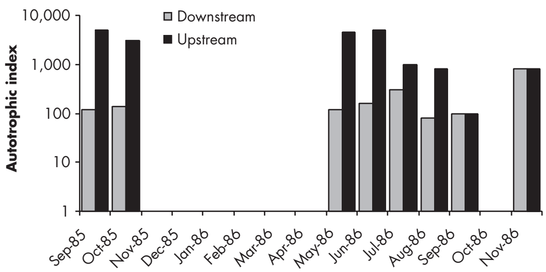

Figure 7.9 Recovery in water quality after improved waste water treatment at an abattoir. The waste water treatment was implemented with progressive reductions in effluent discharged into the river from May 1986. See text for explanation of vertical axis.

Source: Redrawn from Biggs (1989)

In New Zealand there is an ongoing study to improve the water quality in Lake Rotorua in the Central North Island. This is a lake of tremendous importance for tourism and of great cultural importance to the local Māori people. Initially it was thought that the water quality problem could be solved through treating the point source pollution at a sewage treatment plant which received a significant upgrade in 1990. Although this caused a temporary decrease in nutrient loading to the lake the water quality has continued to decline, largely due to agricultural intensification in the lake catchment area. Nitrate-nitrogen levels in the streams feeding into the lake are in the order of 1–2 mg/l but have increased significantly over the past thirty years (White *et al*., 2007). A major concern is that the groundwater levels of nitrate-nitrogen are higher than this, effectively delaying the movement of nutrient to the lake but also making restoration of the lake a very long-term project. Planned action for improving Lake Rotorua water quality include diverting a spring-fed stream away from the lake and buying up intensively farmed land to change the land use to low input forest (White *et al*., 2007). These are expensive options that will take many years to implement and for which it will take even longer to see the results.

SUMMARY

The measurement and management of water quality in a river is an important task within hydrology. To carry this out, a knowledge of the pollution type, pollution source (assuming it is not natural) and pathways leading into the stream are important. Equally, it is important to know the flow regime of any receiving river so that dilution rates can be assessed. There are methods available to control water quality, whether through treatment at point sources (e.g. waste water treatment) or control throughout a catchment using land use planning.

ESSAY QUESTIONS

1 **Explain the Hjulstrom curve and describe its importance for suspended loading in a river.**

2 **Discuss the importance of the BOD_5 test in the assessment of overall water quality for a river.**

3 **Compare and contrast the direct measurement of water quality parameters to the use of proxy measures for the overall assessment of water quality in a river.**

4 **Explain the major causes of enhanced (cultural) eutrophication in a river system and describe the measures that may be taken to prevent it occurring.**

5 **Explain how residence time of water in a catchment can influence the water quality response to land use change.**

FURTHER READING

Chapman, D. (ed.) (1996) *Water quality assessments: a guide to the use of biota, sediments and water in environmental monitoring* (2nd edn). Chapman & Hall, London.
A comprehensive guide to water-quality assessment.

Dodds, W.K. (2002) *Freshwater ecology: concepts and environmental applications*. Academic Press, San Diego.
A comprehensive introduction to water quality impacts on the aquatic environment.

Gray, N.F. (1999) *Water technology: an introduction for environmental scientists and engineers*. Arnold, London.
An introduction to the engineering approach for controlling water pollution.

8

WATER RESOURCE MANAGEMENT IN A CHANGING WORLD

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of water resource management.
- An understanding of the main issues of change that affect hydrology.
- An understanding of how hydrological investigations are carried out to look at issues of change.
- A knowledge of the research literature and main findings in the issues of change.
- A knowledge of case studies looking at change in different regions of the world.

We live in a world that is constantly adjusting to change. This applies from the natural, through to the economic world and is fundamental to the way that we live our lives. The theory of evolution proposes that in order to survive each species on the planet is changing over a long time period (through natural selection) in order to adapt to its ecosystem fully. Equally, economists would say that people and businesses need to adapt and change to stay competitive in a global economy. If, as was argued in the introductory chapter of this book, water is fundamental to all elements of our life on this planet, then we would expect to see hydrology constantly changing to keep up with our changing world. It is perhaps no great surprise to say that hydrology has, and is, changing – but not in all areas. The principle of uniformitarianism states in its most elegant form: 'the present is the key to the past'. Equally, it could be said that the present is the key to the future and we can recognise this with respect to the fundamentals of hydrology. By the end of the twenty-first century people may be living in a different climate from now, their economic lives may be unlike ours, and almost certainly their knowledge of hydrological processes will be greater than at present. However, the hydrological processes will still be operating in the same manner, although may be at differing rates than those that we measure today.

The early chapters of this book have been concerned with hydrological processes and our assessment of them. Our knowledge of the processes will improve, and our methods of measurement and estimation will get better, but the fundamentals will still be the same. In this final chapter several hydrological issues are explored with respect to managing water resources and change that might be expected. The issue of change is explored in a water resource management context: how we respond to changes in patterns of consumption; increasing population pressure and possible changes in climate. The topics discussed here are not exhaustive in covering all issues of change that might be expected in the near future, but they do reflect some of the major concerns. It is meant as an introduction to issues of change and how they affect hydrology; other books cover some of these issues in far more depth (e.g. McDonald and Kay, 1988; Acreman, 2000). The first broad topic of discussion is water resource management, particularly at the local scale. The second topic is the one that dominates the research literature in natural sciences at present: climate change. The third and fourth topics are concerned with the way we treat our environment and the effect this has on water resources: land use change and groundwater depletion. The final topic is urban hydrology – of great concern, with an ever-increasing urban population all around the world.

WATER RESOURCE MANAGEMENT

When the topic of water resource management is discussed it is often difficult to pinpoint exactly what authors mean by the term. Is it concerned with all aspects of the hydrological cycle or only with those of direct concern to humans, particularly water consumption? As soon as the term 'resource' is introduced then it automatically implies a human dimension. Water is a resource because we need it, and there are ways that we can manipulate its provision, therefore water resource management is a very real proposition. If we are going to manage the water environment is it purely for consumption or are there other uses that need protection and management? Over the past hundred years there has been a large-scale rise in the amount of time people in the western world spend at leisure. Leisure activities include sports such as fishing, canoeing and boating, all of which require clean, fully flowing rivers. Thought is now given to the **amenity value** of rivers and lakes (i.e. how useful they are as places of pleasure without necessarily providing a direct economic return). Management of the water environment needs to be designed to maintain and enhance amenity values. Equally we have an obligation to protect the water environment for future generations and for other species that co-exist with the water. Therefore water resource management needs to embrace sustainable development in its good practices. It is clear that water resource management has to embrace all of these issues and at the same time adapt to changing views on what is required of water management.

Almost all of the processes found in the hydrological cycle can be manipulated in some way. Table 8.1 sets out some of these interventions and the implications of their being dealt with by those involved in water resource management. It is immediately apparent from Table 8.1 that the issues go far beyond the river boundary. For example, land use change has a huge importance for water resource management, so that any decisions on land use need to include consultation with water resource managers. It is important that a legislative framework is in place for countries so that this consultation does take place. Likewise for other areas where human intervention may have a significant impact on water resources. The issue of finding the correct management structures and legislation is investigated in the Case Study looking at how water resources have been managed in England over the past forty years (see pp. 153–155). The changing world in this case has been through increasing population pressures, but (probably more importantly) has had to adapt to changing political beliefs.

A key part of water resource management involves water allocation: the amount of water made available to users, including both out of stream users (e.g. irrigation, town water supply) and instream environmental use (e.g. amenity values, supporting aquatic populations). Water allocation in a resource

Table 8.1 Manipulation of hydrological processes of concern to water resource management

Hydrological process	*Human intervention*	*Impact*
Precipitation	Cloud seeding	Increase rainfall (?)
Evaporation	Irrigation	Increase evaporation rates
	Change vegetation cover	Alter transpiration and interception rates
	Change rural to urban	Increase evaporation rates
Storage	Change land use	Alter infiltration rates
	Aquifer storage and recovery (ASR)	Manipulating groundwater storage
	Land drainage	Lowering of local water tables
	Building reservoirs	Increasing storage
Runoff	Change land use	Alter overland flow rates
	Land drainage	Rapid runoff
	River transfer schemes	Alter river flow rates
	Water abstraction	Removing river water and groundwater for human consumption

Case study

CHANGING STRUCTURES OF WATER RESOURCE MANAGEMENT – ENGLAND AS AN EXAMPLE

The major issues of concern for water resource management in England are: water supply; waste disposal; pollution and water quality; and fisheries/aquatic ecosystems management. Other inter-related issues that come into water resource management are flood defence and navigation. Historically it is the first three that have dominated the political agenda in setting up structures to carry out water resource management in England.

History of change

Towards the end of the nineteenth century great municipal pride was taken in the building of reservoirs to supply water to urban centres in England. At the same time many sewage treatment works were built to treat waste. These were built and run by local councils and replaced a previously haphazard system of private water supply and casual disposal of waste. At this stage water resource management resided firmly at the local council level. This system continued until the Water Act of 1973 was passed, a bill that caused a major shake up of water resource management in England and Wales. The major aim of the 1973 act was to introduce holistic water management through administrative boundaries that were governed by river catchments rather than political districts. There was some success in this regard with Regional Water Authorities (RWAs) taking over the water management issues listed above from local councils and other bodies. One of the difficulties with this management structure was the so-called 'poacher–gamekeeper' problem whereby the RWAs were in charge of both waste disposal and pollution control; creating a conflict of interest in water management. Throughout their existence the RWAs operated from a diminishing funding base which led to a lack of investment in waste treatment facilities. It was obvious by the end of the 1980s that a raft of upcoming European Community legislation on water quality would require a huge investment in waste treatment to meet water quality standards. The government at the time decided that this investment was best supplied through the private sector and in 1989 a new Water Act was introduced to privatise the supply of drinking water and wastewater treatment. This has created a set of private water companies with geographic boundaries essentially the same as the RWAs. At the same time a new body, the National Rivers Authority (NRA), was set up to act as a watchdog for water quality. This management structure was still in place in 2002 except that since 1996 the NRA had been subsumed within a larger body, the Environment Agency. The Environment Agency (amongst other duties) monitors river water quality, prosecutes polluters, and issues licenses for water abstraction and treated waste disposal.

How has this change affected water resource management?

The answer to this question can be answered by looking at figures for water abstraction and measured water quality over time.

Figure 8.1 shows the water abstracted for supply in England and Wales from 1961 to 2000. During the period of public control (whether councils or RWAs) there was a steady increase in the amount of water abstracted, apart from a blip in the mid-1970s when there were two particularly dry years. Since privatisation in 1989 there has been a flattening and then decline in the amount of water abstracted. This decline cannot be accounted for by the population which has shown a gradual increase during the same time period (see Figure 8.1). There are two causes of this decline: less water being used for industry due to a decline in the industrial sector (although this has been in decline since the early 1980s), and a drop in the amount of leakage from the supply network. This second factor has been forced upon the water companies by political pressure, particularly following a drought in 1995 and allegations of water supply mismanagement in Yorkshire Water plc. The reduction in leakage has required considerable investment of capital into water supply infrastructure. Overall the water abstracted for public supply is now at the same level as the late 1970s, despite a population rise of nearly 4 million in the corresponding period. The decline has also been achieved despite an increase in the amount of water consumed per household. The United Kingdom has the highest water consumption per capita in Europe and this is rising, a reflection of changing washing habits and an increase in use of dishwashers. This decline in water abstractions is good for the aquatic environment as it allows a more natural river regime and groundwater system to operate with less human intervention.

It is more difficult to ascertain how the changing management structure has affected water quality in England and Wales because the ways of describing water quality have changed with time. Figure 8.2 shows river water quality assessment using three different scales. The figures shown are achieved by sampling water quality over a period of time (normally years) for hundreds of river reaches around the country. During the first

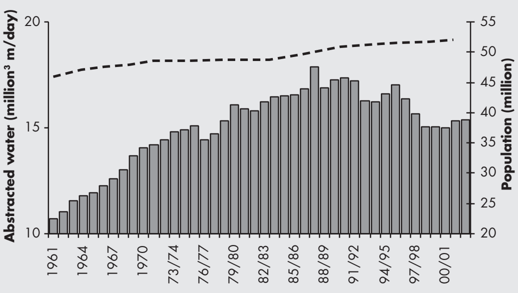

Figure 8.1 Abstracted water for England and Wales 1961–2003 (bar chart) with population for England and Wales 1971–2001 shown as a broken line.

Source: Data from OFWAT and various other sources period, when the control was either local council or RWAs, there was very little change when assessed on a four-point scale. After 1980 the scale was changed to five points and during this period of predominantly RWA control the A grade water quality declined while the percentage of 'fair' water quality river reaches increased. In 1995 the scale was changed again to make it five points and also a differentiation was made between biological and chemical monitoring of water quality (only the chemical results are shown in Figure 8.2). The measurements for 1990 have been recalculated onto the 5-point scale to provide some continuity between assessments. Since 1990 there has been a rise in the two highest categories of water quality at the expense of all the others, a response to the extra investment in wastewater treatment provided by the privatised water companies. The percentage figures for 2000 suggest that the lowest category of water quality has been almost eliminated, although when this is recalculated as river length it shows that there are approximately 162 km of extremely poor quality river reaches (out of a total 40,588 km assessed). The 2005 figures show very little change from 2000.

In summary it has to be said that the biggest impact on the water environment for England and Wales has been the privatisation of supply and wastewater treatment, and the setting up of a separate environmental watchdog organisation, in 1989. Prior to this the water quality remained static or worsened and the amount of water abstracted continued to rise. Although the integration of water management into a holistic structure based around the water catchments (i.e. the RWAs) was a noble idea it made very little difference to the crude measures shown here. Since privatisation the water quality has improved and the total amount of water abstracted has decreased. Fundamentally the reason for this is that the investment in infrastructure has risen dramatically since privatisation. It is probably reasonable to surmise that given the same increase

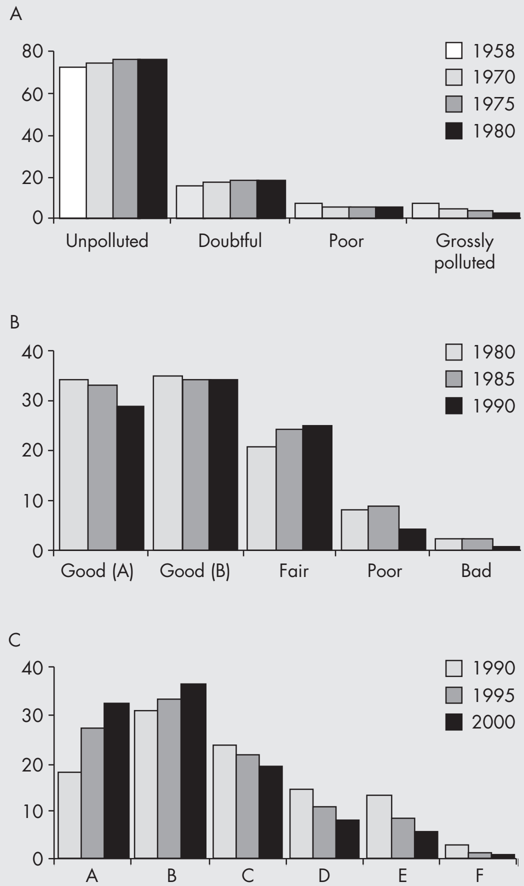

Figure 8.2 Water quality assessment for three periods between 1985 and 2000. Explanation of differing scales is given in the text.

in investment a RWA structure would have seen a similar improvement. It was never likely, in the political climate of the late 1980s, that this investment would have come from the public purse. It has been left to private companies, and more particularly their customers, to pick up the cost of that investment. The example of changing water management structure in England and Wales shows us that this type of change can have significant impacts on the overall hydrology of a region.

management context is about how to ensure fair and equitable distribution of the water resource between groups of stakeholders. In South Africa, legislation introduced in 1998 designated that water for minimum human and ecological needs constitutes an untouchable reserve (Jaspers, 2001). This promotes human usage and instream ecology above other usages (e.g. agriculture, industry). In the USA the way in which 'water rights' are associated with land property rights means that there are many examples where farms have been bought specifically for the associated water right rather than the agricultural value of the land. This is particularly true in western states of the USA like Colorado, where water is definitely a scarce resource. The city of Boulder has steadily acquired agricultural water rights which it has then used for municipal supply. In the 1990s Boulder 'gave back' $12M of water rights to ensure continuous flows in Boulder Creek (Howe, 1996). This is essentially a reallocation of Boulder Creek water in recognition of aesthetic and environmental needs ahead of human usage.

Case study

WATER ALLOCATION IN THREE CONTRASTING COUNTRIES

Demands for water vary according to the climate of an area and the traditional uses of water. This is illustrated in Figure 8.3 which contrasts the uses of abstracted water in New Zealand, the United Kingdom and South Korea. These are three countries of similar size (area) but quite different water demands. New Zealand has a small population (4 million in 2006) with a relatively small industrial sector in an economy dominated by agriculture. This result is that the largest water abstractions are from agriculture, the majority of which is used in the spray irrigation of pasture. The United Kingdom has a much higher population (60 million in 2005) with a large industrial sector compared to agriculture. Irrigation of agricultural land is not common in the UK, hence the small percentage of water used by agriculture compared to industry and household consumption. South Korea has a large population (48 million in 2005) and large industrial economic sector and yet in Figure 8.3 the water abstraction usages are more similar in profile to New Zealand than the United Kingdom. This is because the predominant agricultural practice used, paddy field rice production, has a very high water usage. So although agriculture is not a large part of the South Korean economy there is a high water demand because of the way it is used. The predominance of high-tech industry in South Korea also means that industrial water demand is relatively low compared to what might be expected from heavy industry (e.g. steel production). The large populations of South Korea and the UK result in a much higher percentage of total water allocation going to household supply than in New Zealand. It is estimated globally that although rainfall provides about 90 per cent of water used

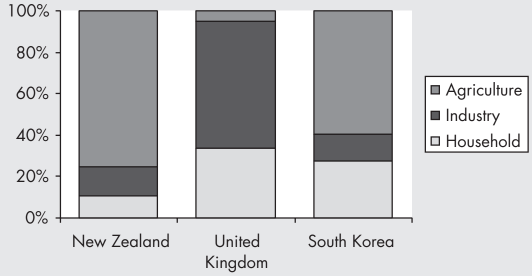

Figure 8.3 Water allocation in three contrasting countries: New Zealand, United Kingdom and South Korea. The *figures* are broad categories of use for water abstracted in each country.

Source: OECD

by crops, 70 per cent of all abstracted water is used in irrigation (UNESCO, 2006). This suggests that the figures on water used for agriculture in the United Kingdom are unusual and South Korea and New Zealand are closer to the norm.

Water allocation demands do not normally stay static. This is illustrated in Figure 8.4, showing the amount of irrigated land in New Zealand from 1965 to 2002. There has been a doubling of irrigated land approximately every twenty years but even within this period there has been irregular development. Between 1985 and 2002 the majority of irrigation development has occurred since 1995. The increase in irrigation has placed significant stresses on the water system and water allocation regime. Improvements in irrigation technology means that a doubling of the irrigated area does not equate to a doubling in water demand, but it still requires an additional quota of either ground or surface water that is not available for other users, especially instream use.

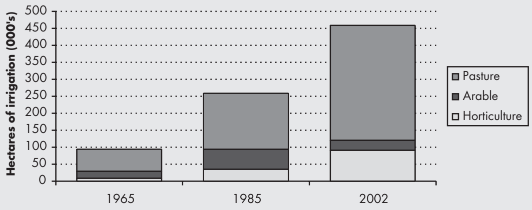

Figure 8.4 Hectares of irrigation in New Zealand from 1965 to 2002.

A key part of water resource management is the involvement of many different sectors of the community in decision-making. This has lead to a different approach to water management that stresses integration between different sectors. There are two key concepts in this area: Integrated Water Resource Management (IWRM) and Integrated Catchment Management (ICM). These are both described in more detail below.

Integrated Water Resource Management (IWRM)

The concepts behind IWRM lie in the so-called 'Dublin Principles'. In January 1992, 500 participants, including government-designated experts from 100 countries and representatives of 80 international, intergovernmental and non-governmental organisations attended the International Conference on Water and the Environment in Dublin, Ireland. The conference adopted what has been termed 'the Dublin Statement' which was taken forward to the Earth Summit Conference in Rio de Janiero later that year.

The Dublin Statement established four guiding principles for managing freshwater resources, namely:

1. Fresh water is a finite and vulnerable resource, essential to sustain life, development and the environment.
2. Water development and management should be based on a participatory approach, involving users, planners and policy makers at all levels.
3. Women play a central part in the provision, management and safeguarding of water.
4. Water has an economic value in all its competing uses and should be recognised as an economic good.

These four principles underlie IWRM, especially the concepts of a participatory approach and that water has an economic value (Solanes, 1998). An economic good, as used in principle four, is defined in economics as: a physical object or service that has value to people and can be sold for a non-negative price in the marketplace. A major implication from principle four is that water is not a gift or a free right to any water user, it needs to be recognised that using water restricts the usage by others and therefore there is a cost involved in the action.

The Global Water Partnership (www.gwpforum.org) is a leading agency in promoting IWRM. It defines IWRM as:

> a process which promotes the coordinated development and management of water, land and related resources in order to maximise the resultant economic and social welfare in an equitable manner without compromising the sustainability of vital ecosystems.
>
> (Global Water Partnership, Technical Advisory Committee, 2004, p. 7)

The emphasis within an IWRM approach to water management is on integration between sectors involved in water resources, including local communities (a participatory approach). Although this is promoted as a new approach to resource management it is in many ways a return to traditional values with recognition of the interconnectedness of hydrology, ecology and land management. If there is a large amount of water from a stream allocated to agriculture then there is less available for town water supply and instream ecology. IWRM is a framework for change that recognises this interconnectedness and builds structures to manage water with this in mind. It is an attempt to move away from structures that promote individual sectors competing against each other for the scarce resource of water and moves towards joint ownership of water resource management.

The types of approaches suggested for use within IWRM are illustrated in Table 8.2. These are instruments for change that the Global Water partnership promotes as being integral to IWRM. While these are by no means the only means of achieving an

Table 8.2 Eight IWRM instruments for change as promoted by the Global Water Partnership

IWRM instrument for change	*Comments and requirements*
Water resources assessment	Understanding what water resources are available and the water needs of communities. Requires measurements of flows, groundwater levels, etc. and water usage (e.g. metering of take).
IWRM plans	Combining development options, resource use and human interaction. Requires inter-sectoral approach.
Demand management	Using water more efficiently. Requires knowledge of where water losses occur (leakage) and plans on how to promote water efficiency.
Social change instruments	Encouraging a water-oriented society. Requires community education on the importance of using water wisely.
Conflict resolution	Managing disputes, ensuring sharing of water. Requires promotion of trust between sectors and robust dispute settlement systems.
Regulatory instruments	Allocation and water use limits. Requires good knowledge about the amount of available resource and how the hydrological system responds to stress (either natural or human-induced).
Economic instruments	Using value and prices for efficiency and equity. Requires good information on water usage and overall water demand.
Information management and exchange	Improving knowledge for better water management. Requires good data-sharing principles (e.g. between flood control and water supply agencies).

Source: GWP (2004)

integrated management of water resources they are a useful starting point.

Integrated Catchment Management (ICM)

Integrated Catchment Management (also sometimes referred to as Integrated Water Basin Management, IWBM) is essentially a subset of IWRM. It aims to promote an integrated approach to water and land management but with two subtle differences:

1 ICM recognises the catchment (or river basin) as the appropriate organising unit for understanding and managing water-related biophysical processes in a context that includes social, economic and political considerations;
2 There is recognition of the spatial context of different management actions and in particular the importance of cumulative effect within a catchment.

By defining a river catchment as the appropriate organising unit for managing biophysical processes there is a recognition that hydrological pathways are important and these provide an appropriate management, as well as biophysical, boundary. Cumulative effect refers to the way in which many small actions may individually have very little impact but when combined the impact may be large. This is true for a river catchment system where individual point discharges of pollution may be small but when accumulated within the river they may be enough to cross an environmental threshold.

Fenemor *et al*. (2006) have defined the word 'integrated' in an ICM context using three different connotations:

- integration between the local community, science and policy so that the community is linked into the planning and execution of both science and policy and scientific research is being carried out in an environment close-linked into policy requirements and vice versa (see Figure 8.5);
- integration between different scientific and technical disciplines to tackle multi-dimensional problems;
- spatial integration throughout a watershed so that the cumulative impact of different actions can be assessed.

Using this type of definition ICM can be seen as a process that can be used to implement IWRM. One of the key principles of ICM and IWRM is community involvement through a participatory approach: making sure that everybody can be involved in resource management, not just a few elite within a single organisation. Another key principle of ICM and IWRM is the idea of change. This ranges from extolling change in management structures to cope with modern resource management pressures to making sure the structures can cope with more inevitable changes in the future.

ICM is promoted by UNESCO and the World Meteorological Organisation (WMO) through the Hydrology for the Environment, Life and Policy (HELP) programme. Further details can be found at www.unesco.org/water/ihp/help.

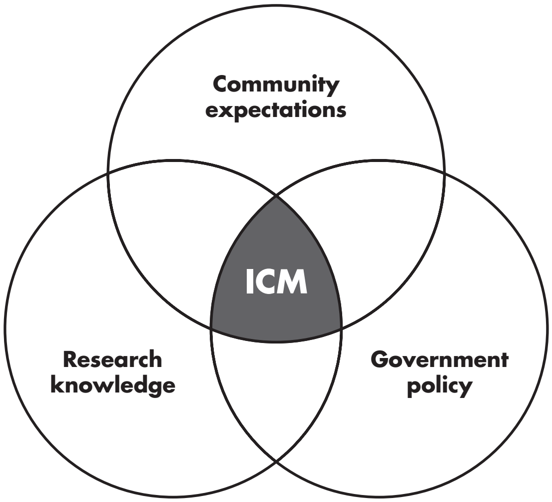

Figure 8.5 The integrating nature of ICM within the context of science, local community and governance.

HYDROLOGY AND CHANGE

In water resource management there is a problem concerning the statistical techniques that we use. In a frequency analysis technique there is an inherent assumption that a storm event with similar antecedent conditions, at any time in the streamflow record, will cause the same size of storm. We assume that the hydrological regime is stationary with time. Under conditions of land use or climate change it is quite possible that these conditions will not be met. This makes it difficult to put much faith in a technique such as frequency analysis when it is known that the hydrological regime has changed during the period of record. These are the types of challenges facing water resource management in an ever-changing world. The following section outlines some of the changes possible and uses case studies to demonstrate the possible effects of those changes.

Climate change

At the start of the twenty-first century climate change is the biggest environmental talking point, dominating the scientific media and research agenda. Any unusual weather patterns are linked to the greenhouse effect and its enhancement by humans. The summer of 2006 in Northern Europe was one of the hottest and driest on record and there was drought. At the same time New Zealand experienced one of the wettest winters on record with record snowfalls to sea level, followed by a wet and cold summer. At various times in the media, both these events were linked to global warming. The difficulty with trying to verify any real link to climate change is that hydrological systems naturally contain a huge amount of variability. The extreme events we are experiencing now may be part of that natural variability, or they may be being pushed to further extremes by climate change. It is unlikely we will know for sure until it is too late to try and do anything about it.

Predictions from the Intergovernmental Panel on Climate Change (IPCC, 2007) suggest that the earth may experience a global surface temperature rise of 0.2°C per decade over the next 100 years. Even if the concentrations of all greenhouse gases and aerosols had been kept constant at year 2000 levels, a further warming of about 0.1°C per decade would be expected. Linked to this prediction are an increase in sea level of 15–95 cm and changes in the temporal and spatial patterns of precipitation. All of these predicted changes will influence the hydrological cycle in some way, but it is difficult to pinpoint exactly how. The IPCC predictions on impacts on water resource matters is shown in Table 8.3. At the very simple level a temperature rise would lead to greater evaporation rates, which in turn puts more water into the atmosphere. This may lead to higher precipitation rates, or at least changes in precipitation patterns. How this impacts the hydrology of an individual river catchment is very difficult to predict. The most common method to make predictions is to take the broad-brush predictions from a global circulation model (often at scale of 1° latitude and longitude per grid square) and downscale it to the local river catchment level. There are several methods used to downscale the data, and Wilby *et al*. (2000) show that the choice of method used can influence the modelling predictions dramatically.

Arora and Boer (2001) have simulated the impacts of possible future climate change on the hydrology of twenty-three major river catchments worldwide. They conclude that in warmer climates there may be a general reduction in annual mean discharge, although as some rivers showed an increase this is not absolute. For mid- to high-latitude rivers Arora and Boer (2001) concluded that there may be big changes in the timing of large runoff events that could be linked to changing seasonal times. This confirms the findings of Middelkoop *et al*. (2001) who predict higher winter discharges on the Rhine (Europe) from 'intensified snow melt and increased winter precipitation'. In a similar vein Wilby and Dettinger (2000) predict higher winter flows for three river basins in the Sierra Nevada (California, USA). These higher winter flows reflect changes in the winter snowpack due to a predicted rise in both precipitation and temperature for the region.

Table 8.3 Predicted impacts of climate change on water resource management area

Water resource area	*Predicted impact*
River flows	There is a high level of confidence that by mid-21st century, annual average river runoff and water availability are projected to increase by 10–40% at high latitudes and in some wet tropical areas, and decrease by 10–30% over some dry regions at mid-latitudes and in the dry tropics, some of which are presently water stressed areas.
Hydrological extremes (floods and droughts)	There is a high level of confidence that drought-affected areas will increase in extent. Heavy precipitation events, which are very likely to increase in frequency, will augment flood risk.
Snow and ice cover	There is a high level of confidence that over the course of the 21st century, water supplies stored in glaciers and snow cover will decline.
Forest production	There is medium confidence that globally, commercial timber productivity will rise modestly with climate change in the short- to medium term, with large regional variability around the global trend. This may affect water yield from forestry-covered catchments.
Coastal flooding	There is very high confidence that many millions more people will be flooded every year due to sea-level rise by the 2080s. The numbers affected will be largest in the mega-deltas of Asia and Africa, while small islands are especially vulnerable.

Source: IPCC (2007)

Arnell and Reynard (1996) used models of river flow to try and predict the effects of differing climate change predictions on the river flows in twenty-one river catchments in Great Britain. Their results suggest a change in the seasonality of flow and also considerable regional variation. Both these changes are by and large driven by differences in precipitation. The North West of England is predicted to become wetter while the South East becomes drier. Overall it is predicted that winters will be wetter and summers drier. This may place a great strain on the water resources for south-east England where by far the greater percentage of people live. In a more recent study Arnell and Reynard (2000) have suggested that flow duration curves are likely to become steeper, reflecting a greater variability in flow. They also predict an increase in flood magnitudes that in the case of the Thames and River Severn have 'a much greater effect than realistic land use change' (Arnell and Reynard, 2000: 21). These changes in river flow regime have important implications for water resource management in the future.

In non-temperate regions of the world predictions vary on climate change. Parry (1990) suggests rainfall in the Sahel region of Africa will stay at current levels or possibly decline by 5–10 per cent. Parry (1990) also suggests a 5–10 per cent increase in rainfall for Australia, although this may have little effect on streamflow when linked with increased evaporation from a 2°C temperature rise. Chiew *et al*. (1995) highlight the large regional variations in predictions of hydrologic change in Australia. The wet tropical regions of north-east

Australia are predicted to have an increase in annual runoff by up to 25 per cent, Tasmania a 10 per cent increase, and a 35 per cent decrease in South Australia, by 2030. The uncertainty of this type of prediction is illustrated by south-east Australia where there are possible runoff changes of ±20 per cent (Chiew *et al*., 1995). Similarly, Kaleris *et al*. (2001) attempted to model the impacts of future climate change on rivers in Greece but concluded that the 'error of the model is significantly larger than climate change impacts' and therefore no firm conclusions could be made. Overall it is difficult to make specific predictions for changes in hydrology as the feedback mechanisms within climate change are not properly understood.

Change in land use

The implications of land use change for hydrology has been an area of intense interest to research hydrologists over the last fifty or more years. Issues of land use change affecting hydrology include increasing urbanisation (see pp. 168–173), changing vegetation cover, land drainage and changing agricultural practices leading to salination.

Vegetation change

In Chapter 3 a Case Study showed the effect that trees have on evaporation and interception rates. This is a hydrological impact of vegetation cover change, a subject that Bosch and Hewlett (1982) review in considerable depth. In general Bosch and Hewlett conclude that the greater the amount of deforestation the larger the subsequent streamflows will be, but the actual amount is dependent on the vegetation type and precipitation amount. This is illustrated by the data in Table 8.4. In the Australian study of Crockford and Richardson (1990) the large range of values are from different

Table 8.4 The amount of interception loss (or similar – see note below) for various canopies as detected in several studies

Canopy cover	*Interception loss*	*Source*
Eucalypt forest (Australia)	5–26% per rainfall event	Crockford and Richardson (1990)
Pine forest (Australia)	6–52% per event	Crockford and Richardson (1990)
Oak stand (Denmark)	15% of summer rainfall	Rasmussen and Rasmussen (1984)
Amazonian rainforest	9% of annual	Lloyd *et al*. (1988)
Sitka spruce (Lancashire, UK)	38% of annual precipitation*	Law (1956)
Sitka spruce (Wales)	30% of annual precipitation*	Kirby *et al*. (1991)
Grassland (Wales)	18% of annual precipitation*	Kirby *et al*. (1991)
Young Douglas fir (New Zealand) (closed canopy)	27% of 7-month summer/ autumn period	Fahey *et al*. (2001)
Mature Douglas fir (NZ)	24% of 7-month summer/ autumn period	Fahey *et al*. (2001)
Young Pinus radiata (NZ) (closed canopy)	19% of 7-month summer/ autumn period	Fahey *et al*. (2001)

Note: *The figures denoted with an asterisk are actually evapotranspiration values rather than absolute interception loss, leading to higher values.

size storms. The high interception losses were experienced during small rainfall events and vice versa. The interception loss from the Amazonian rain forest is remarkably low, reflecting a high rainfall intensity and high humidity levels. Overall there is a high degree of variability in the amount of interception loss that is likely to occur. While it may be possible to say that in general a land use change that has increased tree cover will lead to a water loss, it is not easy to predict by how much that will be.

Fahey and Jackson (1997) conclude that with the loss of forest cover both low flows and peak flows increase. The low flow response is altered primarily through the increase in water infiltrating to groundwater without interception by a forest canopy. The peak flow response is a result of a generally wetter soil and a low interception loss during a storm when there is no forest canopy cover. The time to peak flow may also be affected, with a more sluggish response in a catchment with trees. In a modelling study, Davie (1996) has suggested that any changes in peak flow that result from afforestation are not gradual but highly dependent on the timing of canopy closure.

In Chapter 5 the issue of measurement scale was discussed, and it is particularly pertinent for issues of land use change. There has been considerable debate in the hydrological research literature as to how detectable the effects of deforestation are in large river catchments. Jones and Grant (1996) and Jones (2000) analysed data from a series of paired catchment studies in Oregon, USA and concluded that there was clear evidence of changes in interception rates and peak discharges. Thomas and Megahan (1998) reanalysed the data used by Jones and Grant (1996) and came to the conclusion that although there was clear evidence of changes in peak flows in the small-scale catchment pairs (60–100 km^2) there was no change, or inconclusive evidence for change, in the large catchments (up to 600 km^2). There has followed a series of letters between the authors disputing various aspects of the studies (see *Water Resources Research* volume 37: 175–183). This debate in the research literature mirrors the overall concern in hydrology over the scale issue. There are many processes that we measure at the small hillslope level that may not be important when scaled up to larger catchments.

Land drainage

Land drainage is a common agricultural 'improvement' technique in areas of high rainfall and poor natural drainage. In an area such as the Fens of Cambridgeshire, Norfolk and Lincolnshire in England this has taken the form of drains or canals and an elaborate pumping system, so that the natural wetlands have been drained completely. The result of this has been the utilisation of the area for intensive agricultural production since the drainage took place in the seventeenth and eighteenth centuries. Since that time the land has sunk, due to the removal of water from the peat-based soils, and the area is totally dependent on the pumping network for flood protection. To maintain this network vegetation control and clearance of silt within channels is required, a cost that can be challenged in terms of the overall benefit to the community (Dunderdale and Morris, 1996).

At the smaller scale, land drainage may be undertaken by farmers to improve the drainage of soils. This is a common practice throughout temperate regions and allows soils to remain relatively dry during the winter and early spring. The most common method of achieving this is through a series of tile drains laid across a field that drain directly into a water course (often a ditch). Traditionally tile drains were clay pipes that allowed water to drain into them through the strong hydraulic gradient created by their easy drainage towards the ditch. Modern tile drains are plastic pipes with many small holes to allow water into them. Tile drains are normally laid at about 60 cm depth and should last for at least fifty years or more. To complement the tile drains '**mole drainage**' is carried out. This involves dragging a large, torpedo-shaped metal 'mole' behind a tractor in lines orthogonal to the tile drains. This creates hydrological pathways, at 40–50 cm depth, towards the tile drains. Mole draining may be a regular agricultural activity, sometimes every two to five years in heavy agricultural land (i.e. clay soils). Normal plough depth is around 30 cm, so that the effects of mole draining last beyond a single season.

The aim of tile and mole drainage is to hold less water in a soil. This may have two effects on the overall hydrology. It allows rapid drainage from the field, therefore increasing the flashy response (i.e. rapid rise and fall of hydrograph limbs) in a river. At the same time the lack of soil moisture may lead to greater infiltration levels and hence less overland flow. Spaling (1995) noted that in southern Ontario, Canada, land drainage alters timing and volume of water flow at the field scale, but it is difficult to detect this at the watershed scale. Hiscock *et al*. (2001) analysed sixty years of flow records for three catchments in Norfolk, UK to try and detect any change in the rainfall–runoff relationship during this time. The conclusion of their study was that despite much land drainage during the period of study the rainfall–runoff relationship 'remained essentially unchanged' (Hiscock *et al*., 2001). This lack of change in overall hydrology, despite the known land drainage, may be due to the two hydrological impacts cancelling each other out, or else that the impact of land drainage is small, particularly at the large catchment scale.

Land drainage can be a significant factor in upland areas used for forestry. A common technique in Europe is using the plough and furrow method of drainage. A large plough creates drains in an area, with the seedlings being planted on top of the soil displaced by the plough (i.e. immediately adjacent to the drain but raised above the water table). Like all land drainage this will lower the water table and allow rapid routing of stormflow. A study at a small upland catchment in the north-east of England has shown that land drainage effects are drastic, and only after thirty years of afforestation has the impact lessened (Robinson, 1998). This long recovery time may be a reflection of the harsh environment the trees are growing in; other areas have recovered much faster.

Salination

Salination is an agricultural production problem that results from a build up of salt compounds in the surface soil. Water flowing down a river is almost never 'pure', it will contain dissolved solids in the form of salt compounds. These salt compounds are derived from natural sources such as the weathering of surface minerals and sea spray contained in rainfall. When water evaporates the salts are left behind, something we are familiar with from salt lakes such as in Utah, central Australia, and the Dead Sea in the Middle East. The same process leads to salinity in the oceans.

Salination of soils (often also referred to as salinisation) occurs when there is an excess of salt-rich water that can be evaporated from a soil. The classic situation for this is where river-fed irrigation water is used to boost agricultural production in a hot, dry climate. The evapotranspiration of salt-rich irrigation water leads to salt compounds accumulating in the soil, which in turn may lead to a loss of agricultural production as many plants fail to thrive in a salt-rich environment. Although salination is fundamentally an agronomic problem it is driven by hydrological factors (e.g. water quality and evaporation rates), hence the inclusion in a hydrological textbook.

Salination of soils and water resources have been reported from many places around the world. O'Hara (1997) provides data on waterlogging and subsequent salination in Turkmenistan, the direct result of irrigation. Gupta and Abrol (2000) describe the salinity changes that have occurred in the Indo-Gangetic Plains on the Indian sub-continent following increased rice and wheat production. Flugel (1993) provides data on irrigation return flow leading to salination of a river in the Western Cape Province of South Africa. Prichard *et al*. (1983) report salination of soils from irrigation in California, USA. Irrigation water often has a high total dissolved solids (TDS) load before being used for agricultural production. Postel (1993) suggests that typical values range from 200 to 500 mg/l, where water is considered brackish at levels greater than 300 mg/l. Postel also states that using this type of irrigation

Case study

SALINATION OF WATERWAYS IN AUSTRALIA: A SALINITY PROBLEM FROM LAND USE CHANGE

Salination of surface waters is a huge problem for large areas of Australia. Sadler and Williams (1981) estimate that a third of surface water resources in south-west Western Australia can be defined as brackish (from Williamson *et al*., 1987). In the same region it is estimated that 1.8 million hectares of agricultural land is affected by salinity problems (Nulsen and McConnell, 2000). In an assessment of ten catchments in New South Wales and Victoria (total land area of 35.7 million hectares) it is estimated that 4.1 per cent of the land area is affected by salinity and that this imposes a cost of $122 million (Australian dollars) on agricultural production (Ivey ATP, 2000).

This salinity problem is a steady increase in concentration of salt compounds in rivers, leading to the surface water becoming unusable for public supply or irrigation. In south-west Western Australia salinity levels in soils are typically between 20 and 120 kg/m^2 (Schofield, 1989; Williamson *et al*., 1987, quote a range of 0.2–200 kg/m^2). This is very high and is a result of thousands of years of low rainfall and high evaporation leading to an accumulation of wind-borne sea salt in the soil. The natural vegetation for this area is deep-rooted eucalypt forest and savannah woodland which has a degree of salt tolerance and ability to extract water from deep within the soil. The ability to draw water from deep within a soil maintains a high soil water deficit which is filled by seasonal rainfall (Walker *et al*., 1990). The removal of this native vegetation and replacement with shallow rooted crops (particularly wheat) and pasture has lead to a fundamental change in the hydrology, which in turn has lead to a change in salinity. The replacement vegetation does not use as much water, leading to greater levels of groundwater recharge and rising water tables. This is particularly so for wheat which is largely dormant, or when the ground is fallow, during the wetter winter season. The groundwater is often saline and in addition to this, as the water table rises it takes up the salt stored in soils. Rivers receive more groundwater recharge (with saline water) and the streams increase in salinity. The link between vegetation change and increasing salinity levels was first proposed by Wood (1924) and has since been demonstrated through field studies.

Williamson *et al*. (1987) carried out a catchment based study of vegetation change and increasing salinity in Western Australia. The study monitored salinity and water quantity in four small catchments, two of which were cleared of native vegetation and two kept as controls. The monitoring took place between 1974 and 1983 with the vegetation change occurring at the end of 1976 and start of 1977; a selection of results is shown here. There was a marked change in the hydrological regime (see Figure 8.6), with a large

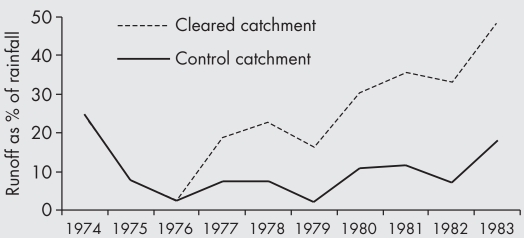

Figure 8.6 Streamflow expressed as a percentage of rainfall for two catchments in south-west Western Australia. The control maintained a natural vegetation while in the other catchment the bush was cleared during 1976/77 and replaced with pasture.

Source: Data from Williamson *et al*. (1987)

increase in the amount of streamflow as a percentage of rainfall received. This reinforces the idea of Wood (1924) that the native vegetation uses more water than the introduced pasture species.

In terms of salinity there was also a marked change although this is not immediately evident from a time series plot (Figure 8.7). The chloride concentration in streamflow is a good indicator of salinity as it is one of the main salts that would be expected to be deposited from sea spray, however it is not the total salinity. In Figure 8.7 there appears to be an increasing difference between chloride concentrations with time. Chloride concentration shows considerable variation between years which is related to variation in rainfall between years. The peaks in salinity correspond to years with high rainfall. To remove this factor Williamson *et al*. (1987) calculated the chloride concentration as a ratio between output (measured in the streamflow) and input (measured in the rainfall). This is shown in Figure 8.8.

When the chloride level is expressed as this output/input ratio (Figure 8.8) it is easy to see a marked difference following the vegetation change. In the years following 1976/77 there is considerably more output of chloride than input (i.e. the ratio is well above a value of 1), a result of the chloride being leached out of the soil. In this manner the chloride concentration in the river is staying at a high level even when there is a low input (i.e. low rainfall). Prior to vegetation change the ratio is approximately even, the chloride inputs and output had reached some type of equilibrium. Given enough time the same would happen again with the new vegetation cover, but first a large store of chloride would be released from the soil. This is a case where the vegetation change has upset the hydrological balance of a catchment, which in turn has implications for water quality.

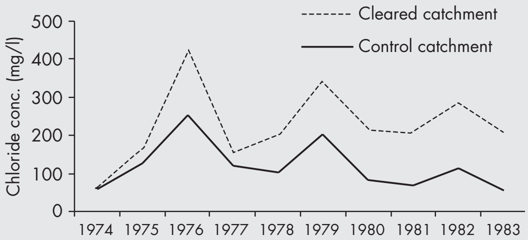

Figure 8.7 Chloride concentrations for two catchments in south-west Western Australia. These are the same two catchments as in *Figure* 8.6. NB World Health Organisation guidelines suggest that drinking water should have chloride concentration of less than 250mg/l.

Source: Data from Williamson *et al*. (1987)

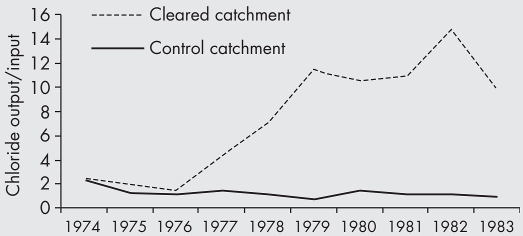

Figure 8.8 Chloride output/input ratio for two catchments in south-west Western Australia. These are the same two catchments as in Figures 8.6 and 8.7. Input has been measured through chloride concentrations in rainfall while output is streamflow.

Source: Data from Williamson *et al*. (1987)

water under a normal irrigation level would add between 2 and 5 tons of salt per hectare per year. The vast majority of this salt is washed out of the soil and continues into a water table or river system; some, though, will be retained to increase salination in the soil.

Groundwater depeletion

In many parts of the world there is heavy reliance on aquifers for provision of water to a population. In England around 30 per cent of reticulated water comes from groundwater, but that rises to closer to

75 per cent in parts of south-east England. The water is extracted from a chalk aquifer that by and large receives a significant recharge during the winter months. Apart from very dry periods (e.g. the early 1990s) there is normally enough recharge to sustain withdrawals. Not all groundwater is recharged so readily. Many aquifers have built up their water reserves over millions of years and receive very little infiltrating rainfall on a year by year basis. Much of the Saudi peninsula in the Middle East is underlain by such an aquifer. The use of this water at high rates may lead to groundwater depletion, a serious long-term problem for water management. The Ogallala aquifer Case Study introduces groundwater depletion problems in the High Plains region of the USA.

Case study

OGALLALA AQUIFER DEPLETION

The Ogallala aquifer (also called the High Plains aquifer) is a huge groundwater reserve underlying an area of approximately 583,000 km^2 in the Great Plains region of the USA. It stretches from South Dakota to Texas and also underlies parts of Nebraska, Wyoming, Colorado, Kansas and New Mexico (see Figure 8.9).

The aquifer formed through erosion from the Rocky Mountains to its west. The porous material deposited from this erosion was filled with water from rivers draining the mountains and crossing the alluvial plains. This has created a water reserve that in places is 300 m deep. A major problem is that now the aquifer is isolated from the Rocky Mountains as a recharge source and has to rely on natural replenishment from local rainfall and infiltration. This is a region that receives around 380–500mm of rainfall per annum and has very high evaporation rates during the summer. The climate is classified as semi-arid.

Ever since Europeans first settled on the Great Plains the Ogallala aquifer has been an important water source for irrigation and drinking water supply. Since the 1940s there has been rapid expansion in the amount of irrigated land in the region (see Figure 8.10) so that in 1990 as much as 95 per cent of water drawn from the aquifer was used for irrigating agricultural land (McGuire and Fischer, 1999). Improving technology has meant that the windmill driven irrigation that was predominant in the 1940s and 1950s has been replaced with pumps capable of extracting vast amounts of water at a rapid rate. The result of this

Figure 8.9 Location of the Ogallala aquifer in the Midwest of the USA.

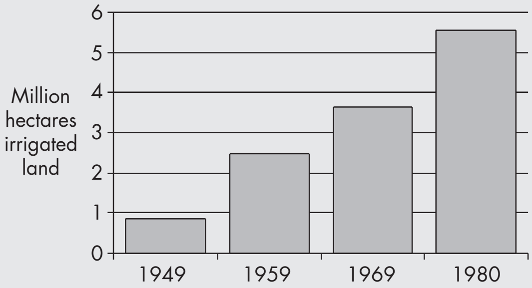

Figure 8.10 Amount of irrigated land using groundwater in the High Plains region.

Source: Data from McGuire and Fischer (1999)

has been drastic declines in water tables, as much as 30 m in parts of Texas, New Mexico and Kansas (McGuire and Fischer, 1999).

There have been various efforts made to try and reduce the depletion of the Ogallala aquifer but it is made difficult by the importance this area has for agricultural production in the USA. Systems of irrigation scheduling have been introduced to make the use of irrigated water more efficient. This involves a close monitoring of soil moisture content so that water is only applied when needed by plants and the actual amount required can be calculated. Another management tool to lessen depletion is changing agricultural production so that water thirsty plants such as cotton are not grown in areas that rely on groundwater for irrigation.

The United States Geological Survey (USGS) have been monitoring changes in water in over 7,000 wells since the late 1980s in order to assess the rate of overall groundwater depletion. Average figures for the period 1980–1996 are shown in Figure 8.11. This shows an overall decline in water tables of around 0.8m but that some regions have shown a rise (e.g. Nebraska 0.7m rise). Although the overall water table has declined there has been a slowing in the rate of decline. This has been attributed to various factors including: a wetter than usual period from 1980 to 1997; more efficient irrigation usage and technology; regulation of groundwater withdrawals and changing commodity prices in agriculture.

It is encouraging that a decline in the rate of water table drops has occurred in the Ogallala aquifer, but these still represent an unsustainable depletion of the groundwater. It is difficult to see how the decline could be halted without a complete change in agricultural production for the region, but this is unlikely to occur until the price of extracting the water is too high to be economically viable. At the moment the region is using an unsustainable management practice that has led to substantial groundwater depletion and is likely to continue into the near future.

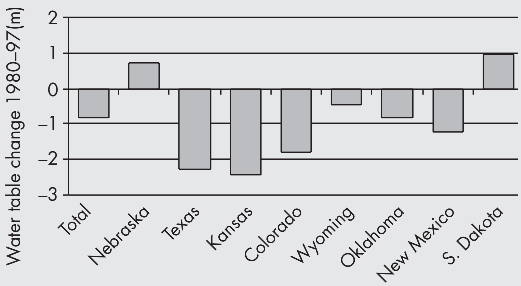

Figure 8.11 Average changes in the water table for states underlying the Ogallala aquifer.

Urbanisation

Many aspects of urban hydrology have already been covered, especially with respect to water quality (see Chapter 7), but the continuing rise in urban population around the world makes it an important issue to consider under the title of change. There is no question that urban expansion has a significant effect on the hydrology of any river draining the area. Initially this may be due to climate alterations affecting parts of the hydrological cycle. The most obvious hydrological impact is on the runoff hydrology, but other areas where urbanisation may have an impact are point source and diffuse pollution affecting water quality, river channelisation to control flooding, increased snow melt from urban areas and river flow changes from sewage treatment.

Urban climate change

In Table 8.5 some of the climatic changes due to urbanisation are expressed as a ratio between the urban and rural environments. This suggests that within a city there is a 15 per cent reduction in the amount of solar radiation reaching a horizontal surface, a factor that will influence the evaporation rate. Studies have also found that the precipitation levels in an urban environment are higher by as much as 10 per cent. Atkinson (1979) detected an increase in summer thunderstorms over London which was attributed to extra convection and condensation nuclei being available. Other factors greatly affected by urbanisation are winter fog (doubled) and winter ultraviolet radiation (reduced by 30 per cent).

Table 8.5 Difference in climatic variables between urban and rural environments

Climatic variable	*Ratio of city: environs*
Solar radiation on horizontal surfaces	0.85
UV radiation: summer	0.95
UV radiation: winter	0.70
Annual mean relative humidity	0.94
Annual mean wind speed	0.75
Speed of extreme wind gusts	0.85
Frequency of calms	1.15
Frequency and amount of cloudiness	1.10
Frequency of fog: summer	1.30
Frequency of fog: winter	2.00
Annual precipitation	1.10
Days with less than 5 mm precipitation	1.10

Source: From Lowry (1967)

Urban runoff change

The changes in climate are relatively minor compared to the impact that impermeable surfaces in the urban environment have on runoff hydrology. Roofs, pavement, roads, parking lots and other impermeable surfaces have extremely low infiltration characteristics, consequently Hortonian overland flow readily occurs. These surfaces are frequently linked to gutters and stormwater drains to remove the runoff rapidly. The result of this is far greater runoff and the time to peak discharge being reduced. Cherkauer (1975) compared two small catchments in Wisconsin, USA. The rural catchment had 94 per cent undeveloped land while the urban catchment had 65 per cent urban coverage. During a large storm in October 1974 (22 mm of rain in five hours) the peak discharge from the urban catchment area was over 250 times that of the rural catchment (Cherkauer, 1975). The storm hydrograph from this event was considerably more flashy for the urban catchment (i.e. it had a shorter, sharper peak on the hydrograph).

Rose and Peters (2001) analysed a long period of streamflow data (1958–96) to detect differences between urbanised and rural catchments near Atlanta, Georgia, USA. The stormflow peaks for large storms were between 30 and 100 per cent larger in the urbanised catchment, with a considerably shorter recession limb of the hydrograph. In contrast to the stormflows, low flows were 25–35 per cent less in the urban catchment, suggesting a lower rainfall infiltration rate. Overall there was no detectable difference in the annual runoff coefficient (runoff as percentage of precipitation) between urban and rural catchments.

Figure 8.12 shows some data from a steadily urbanising catchment (13km^2) in Auckland, New Zealand. There has been a drop in the percentage of baseflow leaving the catchment (the baseflow index – BFI) which could be attributed to declining infiltration to groundwater and therefore less water released during the low flow periods. Care needs to be taken in interpreting a diagram like Figure 8.12 because it is also possible that the decline in BFI was

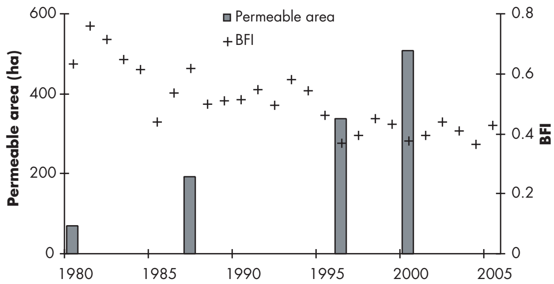

Figure 8.12 Baseflow index (BFI – proportion of annual streamflow as baseflow) with time in a small catchment in Auckland, New Zealand where there has been steady urbanisation. The vertical bars show area of permeable surfaces estimated from aerial photographs at 4 times.

Source: Data courtesy of Auckland Regional Council

caused by an increase in the stormflow and total flow and the actual amount of baseflow has stayed the same. Whichever way, there has been a change in the hydrological regime that can probably be attributed to the rise in permeable surfaces as a result of urbanisation.

Pollution from urban runoff

There is a huge amount of research and literature on the impacts of urbanisation on urban water quality. Davis *et al*. (2001) link the accumulation of heavy metals in river sediments to urban runoff, particularly from roads. Specific sources are tyre wear and vehicle brakes for zinc, and buildings for lead, copper, cadmium and zinc (Davis *et al*., 2001). In Paris, Gromaire-Mertz *et al*. (1999) found high concentrations of heavy metals in runoff from roofs, while street runoff had a high suspended solids and hydrocarbon load. The hydrocarbons are of particular concern, especially the carcinogenic polycyclic aromatic hydrocarbons (PAH) derived from petrol engines. Krein and Schorer (2000) trace PAHs from road runoff into river sediments where they bind onto fine sand and silt particles.

The nature of urban runoff (low infiltration and rapid movement of water) concentrates the pollutants in the first flush of water. Studies have shown that over 80 per cent of pollutant particles are washed into a drainage system within the first 6–10 mm of rain falling (D'Arcy *et al*., 1998), and often from a very small collection area within the urban catchment (Lee and Bang, 2000). This information is important when proposing strategies to deal with the urban pollutant runoff. One of the main methods is to create an artificial wetland within an urban setting so that the initial flush of storm runoff is collected, slowed down, and pollutants can be modified by biological action. Shutes (2001) has a review of artificial wetlands in Hong Kong, Malaysia and England and discussed the role of plants in improving water quality. Scholes *et al*. (1999) showed that two artificial wetlands in London, England were efficient in removing heavy metals and lowering the BOD of urban runoff during storm events. Carapeto and Purchase (2000) reported similar efficiency for the removal of cadmium and lead from urban runoff.

River channelisation

It is a common practice to channelise rivers as they pass through urban areas in an attempt to lessen floods in the urban environment. Frequently, although not always, this will involve straightening a river reach and this has impacts on the streamflow. Simons and Senturk (1977) list some of the hydrological impacts of channel straightening: higher velocities in the channel; increased sediment transport and possible base degradation; increased stormflow stage (height); and deposition of material downstream of the straightening. The impact of urban channelisation is not restricted to the channelised zone itself. The rapid movement of water through a channelised reach will increase the velocity, and may increase the magnitude, of a flood wave travelling downstream. Deposition of sediment downstream from the channelised section may leave the area prone to flooding through a raised river bed.

Case study

THE CHEONGGYECHEON RIVER: A POST-MODERN RIVER RESTORATION PROJECT?

The Cheonggyecheon river (sometimes spelled as Cheong Gye Cheong) is a tributary of the Han river, the main river flowing through Seoul, South Korea. It has been the centre of historical Seoul since the fourteenth century, originally marking a political boundary and being a source of drinking and washing water. The original name for the river was Gaecheoni, which means 'open stream' in Korean; the current name Cheonggyecheon literally means 'stream'. Flooding of the Cheonggyecheon has always been a problem for Seoul, requiring regular dredging and providing a major hazard for the city. In the eighteenth century 200,000 people were used to dredge the stream, widen and strengthen its banks. In the first half of the twentieth century rural depopulation and the migration of people to Seoul led to the river banks becoming overcrowded, with the river acting as a sewer, and there were serious problems with flooding and disease. The Japanese colonial government drew up plans to cover the river and work began in 1937. The covering wasn't completed until 1961, by which time South Korea was an independent state. Between 1967 and 1971 a major expressway was built over the river and the river was essentially forgotten (Figure 8.13). Following concerns in the late 1990s over the safety of the expressway structure it was decided to remove the road and concrete covering and establish the Cheonggyecheon as a riverside recreation area. This was part of an overall Seoul urban revitalisation project promoted by the Mayor Lee Myung-bak.

The term 'restoration' is difficult to apply to the Cheonggyecheon project as the end result is a river far from its original, natural state. However the new Cheonggyecheon fulfils a vital function in modern Seoul: providing open space in a landscape dominated by skyscraper buildings, providing a wildlife corridor, and being a cross between an exhibition area for street sculptures and a civic park (Figure 8.14). The result is a spectacular transformation that had led to much revitalisation of the surrounding district.

Figure 8.13 The Cheonggyecheon expressway covering the river from 1971 to 2003.

Source: Photo courtesy of Seoul Metropolitan City archive

The renovation of the river is estimated to have cost US$360 million, three times an early cost estimation (Lee, 2003). In addition to removing the expressway and concrete cover, the design incorporated new roads, a sewerage and stormwater pipe network and had to allow for flooding. The schematic design is shown in Figure 8.15 which shows how the extra flood capacity has been incorporated beneath the roads, the sewerage and stormwater pipe infrastructure being buried within the stream banks. The water for the river is pumped from the nearby Han river via a water treatment plant. This supplies enough water to maintain an average stream depth of 40 cm. The

Figure 8.14 The Cheonggyecheon river in a 'restored' state, 2006.

Source: Photo courtesy of Seoul Metropolitan City archive

Figure 8.15 Schematic diagram of Cheonggyecheon restoration project, showing infrastructure as well as the river.

Source: Image courtesy of Seoul Metropolitan City archive and Lee (2003).

river contains many artificial weirs and fountains, particularly in its upper reaches, ensuring the water oxygen content stays high. Further downstream there is good enough water quality for invertebrates and fish to establish permanent populations. A survey on the ecology of the stream conducted between March and April 2007 by the Seoul Metropolitan Facilities Management Corporation found thirty species of bird and thirteen species of fish inhabiting the stream environment. In addition to the aesthetic benefit of having a river within a city centre the ambient air temperature adjacent to the stream has dropped by around 2°C and the removal of the expressway has led to a decrease in particulate air pollution. However, the water flowing down the river is around 1.4 m^3/s of predominantly treated drinking water, making it an expensive stream to maintain.

A visit to the Cheonggyecheon river in the evening provides an interesting insight into the way an urban stream can be used as a civic park. The upper reaches of the river are ablaze with multi-coloured neon lights lighting up the stream (see Plate 11), the sculptures along the bank and stepping stones across the stream make it a popular place for people to congregate and walk. The Cheonggyecheon has been transformed from a modernist vision (channelised and covered over) into a post-modern concept where the aesthetic function as a civic park is as important as the biophysical function of the river itself.

More details on the Cheonggyecheon Restoration Project can be found at <http://english.seoul.go.kr/cheonggye>.

Urban snow melt

The influence of urbanisation on snow melt is complicated. Semandi-Davies (1998) suggests that melt intensities are generally increased in an urban area, although shading may reduce melt in some areas. Overall there is a greater volume of water in the early thaw from an urban area when compared to a rural area (Taylor and Roth, 1979; Semandi-Davies, 1998). This may be complicated by snow clearance operations, particularly if the cleared snow is placed in storage areas for later melting (Jones, 1997). In this case the greater mass of snow in a small area will cause a slower melt than if it were distributed throughout the streets.

Waste water input and water extraction

Human intervention in the hydrological regime of a river may be in the form of extraction (for irrigation or potable supply) or additional water from waste water treatment plants. The amount of water discharged from a sewage treatment works into a river may cause a significant alteration to the flow regime. At periods of low flow 44 per cent of the river Trent (a major river draining eastern England) may comprise water derived from waste water effluent (Farrimond, 1980 quoted in Newson, 1995). There are times when the river Lea (a tributary of the Thames, flowing through north-east London) is composed of completely recycled water, which may have been through more than one sewage works. Jones has a startling diagram (1997: 227, Figure 7.9) showing very large diurnal variations in the river Tame that can be attributed to sewage effluent flows from the city of Birmingham, England. In this case the lowest flow occurs at 6 a.m. with a rapid rise in effluent flow that by mid-morning has boosted flow in the river Tame by around 40 per cent (≈2.3 cumecs) (Jones, 1997).

The extra flow that a river derives from sewage effluent may be especially significant if the waste water effluent has been abstracted from another catchment. The water in the river Tame naturally flows into the river Trent before flowing into the North Sea on the east coast of England. A large amount of abstracted water for Birmingham comes from the Elan valley in Wales, a natural tributary of the river Wye which drains into the Bristol Channel on the west coast of England. So in addition to causing diurnal fluctuations in the river Tame downstream of Birmingham, the waste water effluent is part of a water transfer across Great Britain. In a study into low flows in the United Kingdom (i.e. England, Wales, Scotland and Northern Ireland), Gustard *et al*. (1992) identified that 37 per cent of flow gauges were measuring flow regimes subject to artificial influence such as abstraction, effluent discharge or reservoir regulation of the river. This degree of flow alteration is a reflection of the high degree of urbanisation and the high percentage of the population living in an urban environment in the United Kingdom.

SUMMARY

The case studies and different sections in this chapter have shown that there are many aspects of change in hydrology to be considered. Equally there are different management structures and principles that can be used to manage the change. In order to understand and make predictions concerning change it is essential to understand the fundamentals of hydrology: how processes operate in time and space; how to measure and estimate the rates of flux for those processes; and how to analyse the resultant data. The fundamental processes do not change; it is their rates of flux in different locations that alter. It is fundamentally important that hydrology as a science is investigating these rates of change, and finding new ways of looking at the scales of change in the next 100 years.

ESSAY QUESTIONS

1 **Discuss how well the principles of Integrated Water Resource Management are applied to the management of a catchment near you.**

2 Explain the way that human-induced climate change may affect the hydrological regime for a region.

3 Assess the role of land use change as a major variable in forcing change in the hydrological regime for a region near you.

4 Compare and contrast the impact of urbanisation to the impact of land use change on general hydrology within the country where you live.

5 Discuss the major issues facing water resource managers over the next fifty years in a specified geographical region.

FURTHER READING

Acreman, M. (ed.) (2000) *The hydrology of the UK, a study of change*. Routledge, London.
Chapters by different authors looking at change in the UK.

Global Water Partnership (2004) Catalyzing change: a handbook for developing integrated water resources management (IWRM) and water efficiency strategies. Published by GWP.
A handbook on practical steps to achieving IWRM.

Intergovernmental Panel on Climate Change (IPCC) (2007) *Climate change 2007*. Cambridge University Press, Cambridge.
Various reports are published by the IPCC, summaries of which can be found at http://www.ipcc.ch

GLOSSARY

acid rain Acidic rain that forms in the atmosphere when industrial gas emissions (especially sulphur dioxide and nitrogen oxides) combine with precipitating water (NB rain is naturally acidic – acid rain has an enhanced acidity from industrial emissions).

actual evaporation Evaporation which occurs at a rate controlled by the available water (e.g. plant transpiration may be restricted by low soil moisture).

advective energy energy that originates from elsewhere (another region that may be hundreds or thousands of kilometres away) and has been transported to a region (frequently in the form of latent heat) where it becomes available energy.

aerodynamic resistance A term to account for the way that the water evaporating off a surface mixes with a potentially drier atmosphere above it through turbulent mixing. It is a measure of aerodynamic roughness. NB This is directly inverse to aerodynamic conductance.

albedo The reflectivity of a surface (a unit percentage).

alkalinity A measure of the capacity to absorb hydrogen ions without a change in pH (Viessman and Hammer, 1998). This is influenced by the concentration of hydroxide, bicarbonate or carbonate ions.

amenity value A term used to denote how useful an area (e.g. a stretch of river) is for recreation and other purposes.

anemometer Instrument for measuring wind speed.

annual maximum series River flow data used in flood frequency analysis. It takes the highest flow in every year of the period.

aquifer A layer of unconsolidated or consolidated rock that is able to transmit and store enough water for extraction. A confined aquifer has restricted flow above and below it while an unconfined aquifer has no upper limit.

aquifer storage and recovery (ASR) A water resource management technique involving the addition of surface water into an aquifer for storage to be recovered later.

aquifuge A totally impermeable rock formation.

aquitard A geological formation that transmits water at a much slower rate than the aquifer (similar to aquifuge).

areal rainfall The average rainfall for an area (often a catchment in hydrology) calculated from several different point measurements.

artesian water or well Water that flows directly to the surface from a confined aquifer (i.e. it does not require extraction from the ground via a pump). The water in aquifer is under pressure so it is able to reach the surface of a well.

AVHRR (Advanced Very High Resolution Radiometer) A North American Space Agency (NASA) satellite used mainly for atmospheric interpretation.

bankfull discharge The amount of water flowing down a river when it is full to the top of its banks.

baseflow The portion of streamflow that is not attributed to storm precipitation (i.e. it flows regardless of the daily variation in rainfall). Sometimes also referred to as slowflow.

Bergeron process The process of raindrop growth through a strong water vapour gradient between ice crystals and small water droplets.

biochemical oxygen demand (BOD) A measure of the oxygen required by bacteria and other micro-organisms to break down organic matter in a water sample. A strong indicator of the level of organic pollution in a river.

Bowen ratio The ratio of sensible heat to latent heat. This is sometimes used within a method to measure evaporation from a surface.

Boyle's law A law of physics relating pressure (P), temperature (T), volume (V) and concentration of molecules (n) in gases.

canopy storage capacity The volume of water that can be held in the canopy before water starts dripping as indirect throughfall.

capillary forces The forces holding back soil water so that it does not drain completely through a soil under gravity. The primary cause of capillary forces is surface tension between the water and soil surfaces.

catchment The area of land from which water flows towards a river and then from that river to the sea. Also known as a river basin.

channel flow Water flowing within a channel. A general term for streamflow or riverflow.

channelisation The confinement of a river into a permanent, rigid, channel structure. This often occurs as part of urbanisation and flood protection.

cloud seeding The artificial generation of precipitation (normally rainfall) through provision of extra condensation nuclei within a cloud.

condensation The movement of water from a gaseous state into a liquid state; the opposite of evaporation.

condensation nuclei Minute particles present in the atmosphere upon which the water or ice droplets form.

convective precipitation Precipitation caused by heating from the earth's surface (leading to uplift of a moist air body).

covalent bonding A form of molecular bonding where electrons are shared between two atoms in the molecule. This is the strongest form of chemical bond and exists within a water molecule.

cyclonic precipitation Precipitation caused by a low-pressure weather system where the air is constantly being forced upwards.

dewfall (or dew) Water that condenses from the atmosphere (upon cooling) onto a surface (frequently vegetation).

dilution gauging A technique to measure streamflow based on the dilution of a tracer by the water in the stream.

discharge In hydrology discharge is frequently used to denote the amount of water flowing down a river/stream with time (units m^3/s, called cumecs).

effective rainfall The rainfall that produces stormflow (i.e. is not absorbed by soil). This is a term used in the derivation and implementation of the unit hydrograph.

eutrophication A term used to describe the addition of nutrients to an aquatic ecosystem that leads to an increase in net primary productivity. The term 'cultural eutrophication' is sometimes used to indicate the enhanced addition of nutrients through human activity. This may lead to problems with excess weed and algal growth in a river.

evaporation The movement of water from a liquid to a gaseous form (i.e. water vapour) and dispersal into the atmosphere.

evaporation pan A large vessel of water, with a measuring instrument or weighing device underneath that allows you to record how much water is lost through evaporation over a time period.

evapotranspiration A combination of direct evaporation from soil/water and transpiration from plants. The term recognises the fact that much of the earth's surface is a mixture of vegetation cover and bare soil.

field capacity The actual maximum water content that a soil can hold under normal field conditions. This is often less than the saturated water content as the water does not fill all the pore space and gravity drains large pores very quickly.

flash flood A flood event that occurs as a result of extremely intense rainfall causing a rapid rise in water levels in a stream. This is common in arid and semi-arid regions.

flood An inundation of land adjacent to a river caused by a period of abnormally large discharge or sea encroachment on the land.

flood frequency analysis A technique to investigate the magnitude–frequency relationship for floods in a particular river. This is based on historical hydrograph records.

flow duration curve A graphical description of the percentage of time a certain discharge is exceeded for a particular river.

flux The rate of flow of some quantity (e.g. the rate of flow of water as evaporation is referred to as an evaporative flux).

frequency–magnitude The relationship between how often a particular event (e.g. flood) occurs, and how large the event is. In hydrology it is common to study low frequency–high magnitude events (e.g. large floods do not happen very often).

Geographic Information Systems (GIS) A computer program which is able to store, manipulate and display spatial digital data over an area (e.g. maps).

geomorphology The study of landforms and how they have evolved.

gravimetric soil moisture content The ratio of the weight of water in a soil to the overall weight of a soil.

groundwater Water held in the saturated zone beneath a water table. The area of groundwater is also referred to as water in the phreatic zone.

groundwater flow Water which moves down a hydraulic gradient in the saturated (phreatic) zone.

hillslope hydrology The study of hydrological processes operating at the hillslope scale.

Hjulstrom curve The relationship between water velocity and sediment erosion and deposition.

Hortonian overland flow See **infiltration excess overland flow**.

hydraulic conductivity A measure of the ability of a porous medium to transmit water. This is a flux term with units of metres per second. The hydraulic conductivity of a soil is highly dependent on water content.

hydraulic radius The wetted perimeter of a river divided by the cross-sectional area.

hydrogen bonding Bonding between atoms or molecules caused by the electrical attraction between a negative and positive ion. This type of bonding exists between water molecules.

hydrograph A continuous record of streamflow.

hydrograph separation The splitting of a hydrograph into stormflow and baseflow.

hydrological cycle A conceptual model of how water moves around between the earth and atmosphere in different states as a gas, liquid or solid. This can be at the global or catchment scale.

hydrology 'The science or study of' ('logy' from Latin *logia*) and 'water' ('hydro' from Greek *hudor*). Modern hydrology is concerned with the distribution of fresh water on the surface of the earth and its movement over and beneath the surface, and through the atmosphere.

hydrometry The science of streamflow measurement.

hydrophobicity and **hydrophobic soils** The ability of some soils to rapidly swell upon contact with water so that the initial infiltration rate is low. In this case the water will run over the surface as infiltration excess overland flow.

hypsometric method A method for estimating areal rainfall based on the topography of the area (e.g. a catchment).

hysteresis The difference in soil suction at a given water content dependent on whether the soil is being wetted or dried (see soil moisture characteristic curve).

infiltration capacity The rate of infiltration of water into a soil when a soil is fully saturated (i.e. at full capacity of water).

infiltration excess overland flow Overland flow that occurs when the rainfall rate exceeds the infiltration rate for a soil. Also referred to as Hortonian overland flow.

infiltration rate How much water enters a soil during a certain time interval.

infiltrometer An instrument to measure the infiltration rate and infiltration capacity for a soil.

instream flow assessment A combination of hydrology and aquatic ecology used to assess how much water, and the flow regime, that is required by particular aquatic fauna in a river or stream.

integrated catchment management (ICM) A form of integrated water resource management (IWRM) that promotes the river catchment as the appropriate organising unit for understanding and managing water-related biophysical processes in a context that includes social, economic and political considerations. Also sometimes referred to as Integrated Water Basin Management – IWBM.

integrated water resource management (IWRM) A water resource management paradigm that promotes the coordinated development and management of water, land and related resources in order to maximise the resultant economic and social welfare in an equitable manner without compromising the sustainability of vital ecosystems.

interception The interception of precipitation above the earth's surface. This may be by a vegetation canopy or buildings. Some of this intercepted water may be evaporated; referred to as interception loss.

isohyetal method A method for estimating areal rainfall based on the known distribution of rainfall within the area (e.g. a catchment).

jökulhlaup The flood resulting from an ice-dam burst.

kriging A spatial statistics technique that identifies the similarity between adjacent and further afield point measurements. This can be used to interpolate an average surface from a series of point measurements.

LANDSAT (LAND SATellite) A series of satellites launched by the North American Space Agency (NASA) to study the earth's surface.

latent heat The energy required to produce a phase change from ice to liquid water, or liquid water to water vapour. When water moves from liquid to gas this is a negative flux (i.e. energy is lost), whereas the opposite phase change (gas to liquid) produces a positive heat flux.

lateral flow See **throughflow**.

low flow A period of extreme low flow in a river hydrograph (e.g. summer or dry season river flows).

low flow frequency analysis A technique to investigate the magnitude–frequency relationship for low flows in a particular river. This is based on historical hydrograph records.

lysimeter A device for collecting water from the pore spaces of soils and for determining the soluble constituents removed in the drainage. In evaporation studies a lysimeter is a cylinder filled with soil and plants used to measure evaporation from a vegetated surface. This can be done either as a weight loss calculation or through solving some form of the water balance equation.

macropores Large pores within a soil matrix, typically with a diameter greater than 3 mm.

model A representation of the hydrological processes operating within an area (usually a catchment). This is usually used to mean a numerical model, which simulates the flow in a river, based on mathematical representations of hydrological processes.

mole drainage An agricultural technique involving the provision of rapid subsurface drainage routes within an agricultural field.

net radiation The total electromagnetic radiation (in all wavelengths) received at a point. This includes direct solar radiation and re-radiation from the earth's surface.

neutron probe An instrument to estimate the soil water content using a radioactive source of fast neutrons.

open water evaporation The evaporation that occurs above a body of water such as a lake, stream or the oceans.

orographic precipitation Precipitation caused by an air mass being forced to rise over an obstruction such as a mountain range.

overland flow Water which runs across the surface of the land before reaching a stream. This is one form (but not the only form) of runoff.

oxygen sag curve The downstream dip in dissolved oxygen content that can be found after the addition of organic pollution.

partial areas concept The idea that only certain parts of a catchment area contribute overland flow to stormflow; compare to the variable source areas concept.

partial duration series River flow data used in flood frequency analysis. It takes the highest flow peaks from the period of record irrespective of the year in which it occurs (compare with **annual maximum series**).

peakflow See **stormflow**.

perched water table Area where the water table is held above a regional water table, usually due to small impermeable lenses in the soil or geological formation.

pH The concentration of hydrogen ions within a water sample. A measure of water acidity on an inverse logarithmic scale.

phreatic zone The area beneath a water table (i.e. groundwater).

piezometer A tube with holes at the base that is placed at depth within a soil or rock mantle to measure the water pressure at a set location.

pipeflow The rapid movement of water through a hillslope in a series of linked pipes. (NB these can be naturally occurring.)

porosity The percentage of pore space (i.e. air) within a dry soil.

potential evaporation Evaporation which occurs over the land's surface if the water supply is unrestricted.

precipitation In hydrology this is the movement of water from the atmosphere to the earth's surface. This can occur as rain, hail, sleet or snowfall.

quickflow See **stormflow**.

rainfall Precipitation in a liquid form. The usual expression of rainfall is as a vertical depth of water (e.g. mm or inches).

rainfall intensity The rate at which rainfall occurs. A depth of rainfall per unit time, most commonly mm per hour.

rain gauge An instrument for measuring the amount of rainfall at a point for a period of time. Standard rain gauges are measured over a day; continuous rainfall measurement can be provided by special rain gauges such as the tipping-bucket gauge.

rain shadow effect An uneven distribution of rainfall caused by a large high landmass (e.g. a mountain range). On the downwind side of the mountain range there is often less rainfall (i.e. the mountain casts a rain shadow).

rating curve The relationship between river stage (height) and discharge.

recession limb (of hydrograph) The period after a peak of stormflow where the streamflow values gradually recede.

relative humidity How close to fully saturated the atmosphere is (a percentage – 100 per cent is fully saturated for the current temperature).

rising limb (of hydrograph) The start of a stormflow peak.

river A large natural stream of water flowing over the surface and normally contained within a river channel.

river basin The area of land from which water flows towards a river and then in that river to the sea. Also known as the river catchment.

roughness coefficient A term used in equations such as Chezy and Manning's to estimate the degree that water is slowed down by friction along the bed surface.

runoff The movement of liquid water above and below the surface of the earth prior to reaching a stream or river.

salination The build up of salts in a soil or water body.

satellite remote sensing The interpretation of ground (or atmospheric) characteristics based on measurements of radiation from the earth/atmosphere. The radiation measurements are received on satellite-based sensors.

saturated overland flow Overland flow that occurs when a soil is completely saturated.

saturated water content The maximum amount of water that the soil can hold. It is equivalent to the soil porosity, which assumes that the water fills all the pore space within a soil.

saturation vapour pressure The maximum vapour pressure possible (i.e. the vapour pressure exerted when a parcel of air is fully saturated). The saturation point of an air parcel is temperature-dependent and hence so is the saturation vapour pressure.

sensible heat The heat which can be sensed or felt. This is most easily understood as the heat we feel as warmth. The sensible heat flux is the rate of flow of that sensible heat.

snowfall Precipitation in a solid form. For hydrology it is common to express the snowfall as a vertical depth of liquid (i.e. melted) water.

snow pillow An instrument used to measure the depth of snow accumulating above a certain point.

soil heat flux Heat released from the soil having been previously stored within the soil.

soil moisture characteristic curve A measured curve describing the relationship between the capillary forces and soil moisture content. This is also called the suction moisture curve.

soil moisture deficit The amount of water required to fill the soil up to field capacity.

soil moisture tension See **soil suction**.

soil suction A measure of the strength of the capillary forces. This is also called the moisture tension or soil water tension. A dry soil exerts a high soil suction.

soil water Water in the unsaturated zone occurring above a water table. This is also referred to as water in the vadose zone.

specific heat capacity The amount of energy required to raise the temperature of a substance by a single degree.

SPOT French satellite to study the earth's surface.

stage In hydrology this term is used to mean the water level height of a river.

stemflow Rainfall that is intercepted by stems and branches, and flows down the tree trunk into the soil.

stomatal or canopy resistance The restriction a plant places on its transpiration rate through opening and closing stomata in the leaves.

storage A term in the water balance equation to account for water that is not a flux or is very slow moving. This may include snow and ice, groundwater and lakes.

storm duration The length of time between rainfall starting and ending within a storm.

stormflow The portion of streamflow (normally seen in a hydrograph) that can be attributed to storm precipitation. Sometimes also referred to as quickflow or peakflow.

stream A small river.

streamflow Water flowing within a stream channel (or river flow for a larger body of water). Often referred to as discharge.

suction moisture curve See **soil moisture characteristic curve**.

Synthetic Aperture Radar (SAR) A remote sensing technique that uses radar properties, usually of microwaves.

synthetic unit hydrograph A unit hydrograph derived from knowledge of catchment characteristics rather than historical hydrograph records.

tensiometer An instrument used to measure the soil moisture tension.

Thiessen's polygons A method of estimating average rainfall for an area based on the spatial distribution of rain gauges.

throughfall The precipitation that falls to the ground either directly (through gaps in the canopy), or indirectly (having dripped off leaves, stems or branches).

throughflow Water which runs to a stream through the soils. This is frequently within the unsaturated (vadose) zone. This is one form of runoff. Sometimes referred to as lateral flow.

time domain reflectometry (TDR) A method to estimate the soil water content based on the interference of propagated electromagnetic waves due to water content.

total dissolved solids (TDS) The amount of solids dissolved within a water sample. This is closely related to the electrical conductivity of a water sample.

total suspended solids (TSS) The amount of solids suspended within a water sample. This is closely related to the turbidity of a water sample.

transpiration The movement of liquid water from a plant leaf to water vapour in the atmosphere. Plants carry out transpiration as part of the photosynthetic process.

turbidity The cloudiness of a water sample.

ultrasonic flow gauge An instrument that measures stream discharge based on the alteration to a propagated wave over a known cross section.

unit hydrograph A model of stormflow in a particular catchment used to predict possible future storm impacts. It is derived from historical hydrograph records.

vadose zone Area between the water table and the earth surface. The soil/rock is normally partially saturated.

vapour pressure Pressure exerted within the parcel of air by having the water vapour present within it. The more water vapour is present the greater the vapour pressure.

vapour pressure deficit The difference between actual vapour pressure and saturation vapour pressure.

variable source areas concept The idea that only certain parts of a catchment area contribute overland flow to stormflow and that these vary in space and time; compare to the partial areas concept.

velocity–area method A technique to measure instantaneous streamflow through measuring the cross-sectional area and the velocity through the cross section.

volumetric soil moisture content The ratio of the volume of water in a soil to the overall volume of a soil.

water balance equation A mathematical description of the hydrological processes operating within a given timeframe. Normally includes precipitation, runoff, evaporation and change in storage.

water table The surface that differentiates between fully saturated and partially saturated soil/rock.

water vapour Water in a gaseous form.

well A tube with permeable sides all the way up so that water can enter or exit from anywhere up the column. Wells are commonly used for water extraction and monitoring the water table in unconfined aquifers.

wetted perimeter The total perimeter of a cross section in a river.

wilting point The soil water content when plants start to die back (wilt).

zero plane displacement The height within a canopy at which wind speed drops to zero.

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INDEX

Kuliah Hidrologi
Kelas S1-REG FTSP

DEBIT ALIRAN PERMUKAAN

Dosen :

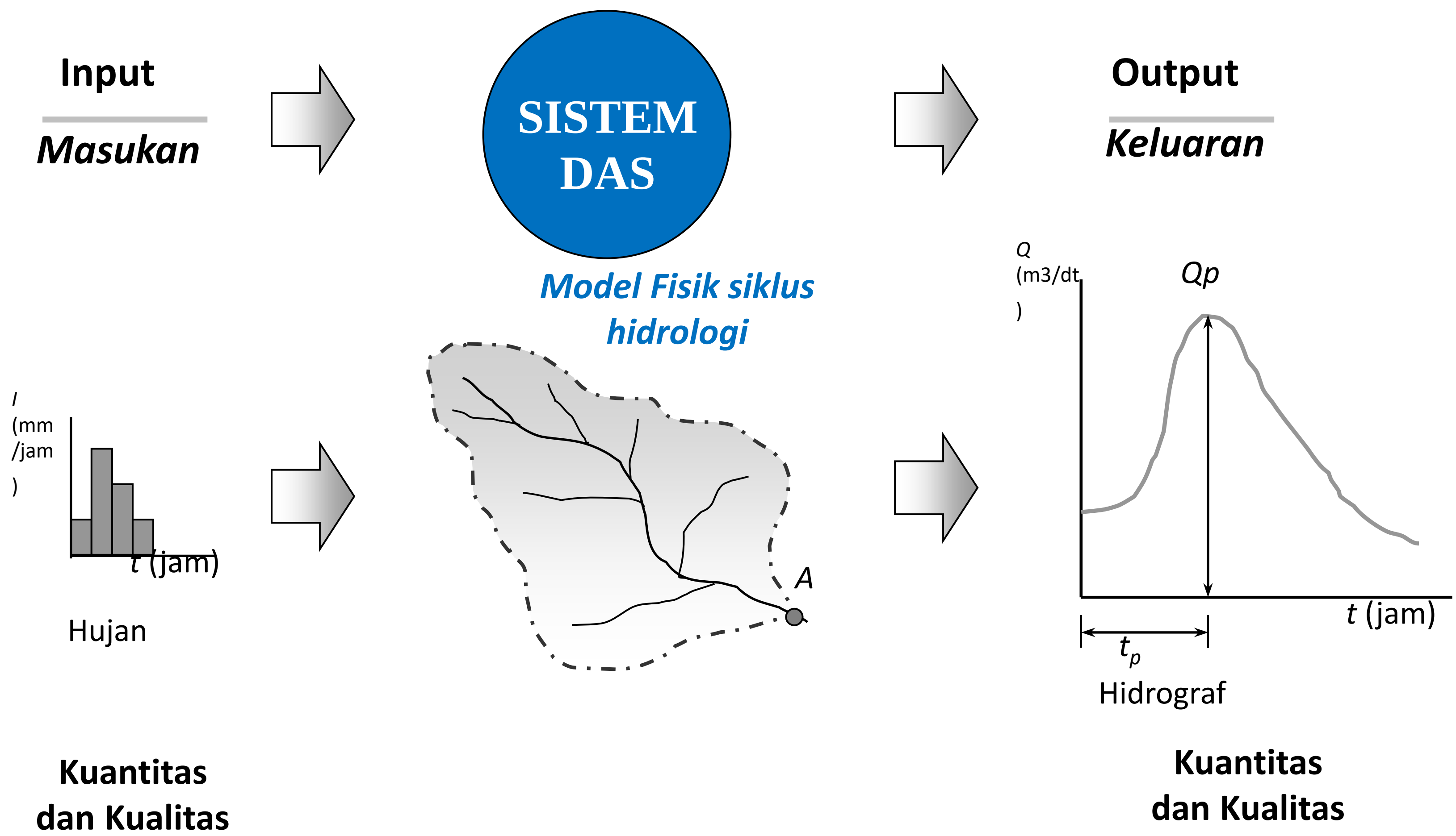
Muhamad Komarudin S.Si., M.Si

Ir. Rahardjo Samiono M.T.

CONTOH APLIKASI HIDROLOGI DALAM BIDANG SIPIL

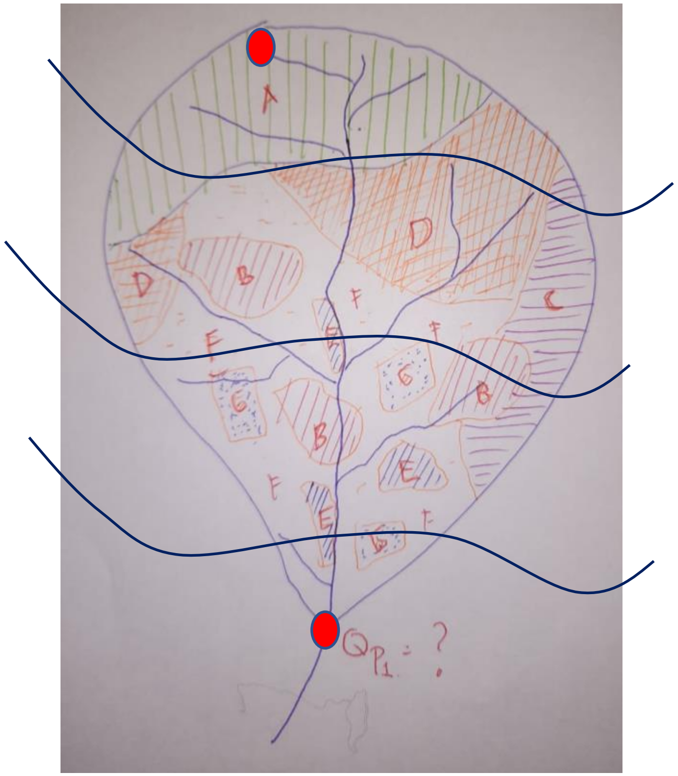
- Dalam Proyek pengembangan sumberdaya air, perkiraan volume air adalah hal yang sangat penting mengingat hal tersebut merupakan dasar perencanaan dan pengoperasian sistem sumberdaya air (Limantara, 2010).
- Proyek sumberdaya air : sistem drainase, supply air untuk irigasi, industry, air minum, pengendalian banjir, PLTA, pengelolaan DAS, penggelontoran, tambak dan lain-lain
- Dalam proyek pengendalian banjir diperlukan nilai ekstrem tebal hujan atau besaran debit (high Flow) dan untuk itu diperlukan Analisa banjir rancangan (design flood).
- Untuk supply (penyediaan) air diperlukan harga ekstrim rendah dari hujan atau debit (low flow) dan untuk keperluan ini dilakukan Analisa debit andalan (dependable discharge)

Skema sistem daerah aliran sungai



Karateristik hidrologi suatu daerah atau Daerah Aliran Sungai (DAS) sangat bergantung pada kondisi geografi dan geologi daerah tersebut.

Perhitungan Debit Puncak Rencana dengan Metode Rasional



Perhitungan besarnya debit banjir rencana dengan metode rasional menggunakan rumus sebagai berikut :

$$Q_t = 0{,}278 \times C \times I_T \times A$$

Keterangan :

Q_t = debit banjir (m^3/dtk)

C = koefisien pengaliran

I_T = intensitas curah hujan dengan periode ulang T tahun (mm/jam)

A = luas areal (km^2)

PEMBAHASAN

Tipe Aliran

Berdasarkan perubahan kedalaman dan/atau kecepatan mengikuti fungsi waktu, maka aliran dibedakan menjadi aliran permanen (*steady*) dan tidak permanen (*Unsteady*).

Berdasarkan fungsi ruang, maka aliran dibedakan menjadi aliran seragam (*Uniform*) dan tidak seragam (*Non Uniform*)

Kriteria Aliran Tetap (*Steady Flow*)

- Perubahan Volume terhadap waktu tetap $\partial Q/\partial t = 0$
- Perubahan Kedalaman terhadap waktu tetap $\partial h/\partial t = 0$
- Perubahan Kecepatan terhadap waktu tetap $\partial v/\partial t = 0$

Kriteria Aliran Tidak Tetap (*Unsteady Flow*)

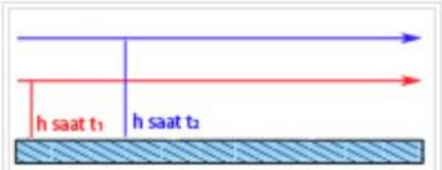
- Perubahan Volume terhadap waktu tidak tetap $\partial Q/\partial t \neq 0$
- Perubahan Kedalaman terhadap waktu tidak tetap $\partial h/\partial t \neq 0$
- Perubahan Kecepatan terhadap waktu tidak tetap $\partial v/\partial t \neq 0$

Kriteria Seragam (*Uniform Flow*)

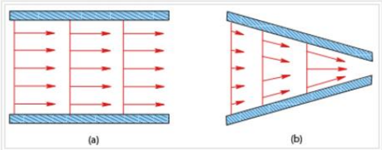
- Besar dan arah kecepatan tetap terhadap jarak $\partial Q/\partial s = 0$
- Aliran pada pipa dengan penampang sama $\partial v/\partial s = 0$
- Variabel Fluida lainnya juga tetap $\partial h/\partial z = 0$

Kriteria Tidak Seragam (*Non Uniform Flow*)

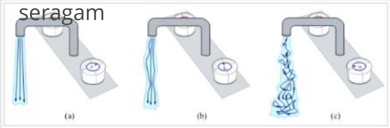
- Besar dan arah kecepatan tidak tetap terhadap jarak $\partial Q/\partial s \neq 0$
- Aliran pada pipa dengan penampang tidak sama $\partial v/\partial s \neq 0$
- Variabel Fluida lainnya juga tidak tetap $\partial h/\partial z \neq 0$



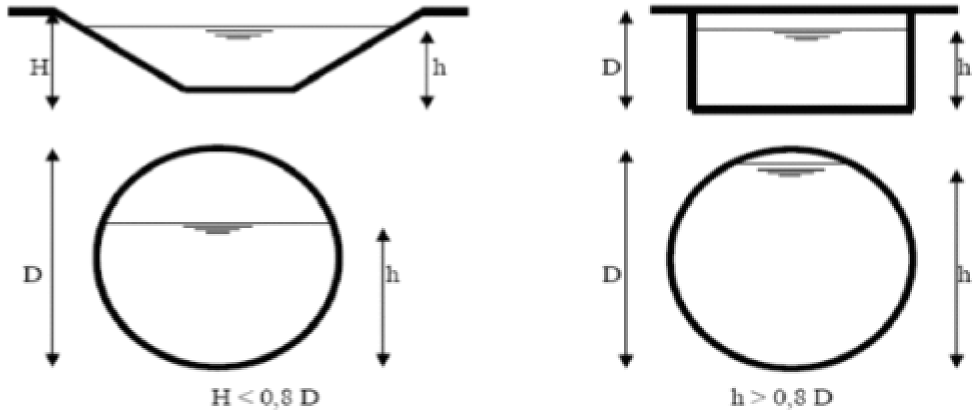
Aliran tak tunak atau aliran tidak permanen (*impermanent flow*) adalah kondisi dimana komponen aliran berubah terhadap waktu. Contoh aliran di saluran/sungai pada kondisi ada perubahan aliran (ada hujan, ada banjir, dll) atau aliran yang dipengaruhi muka air pasang-surut (muara sungai di laut) .



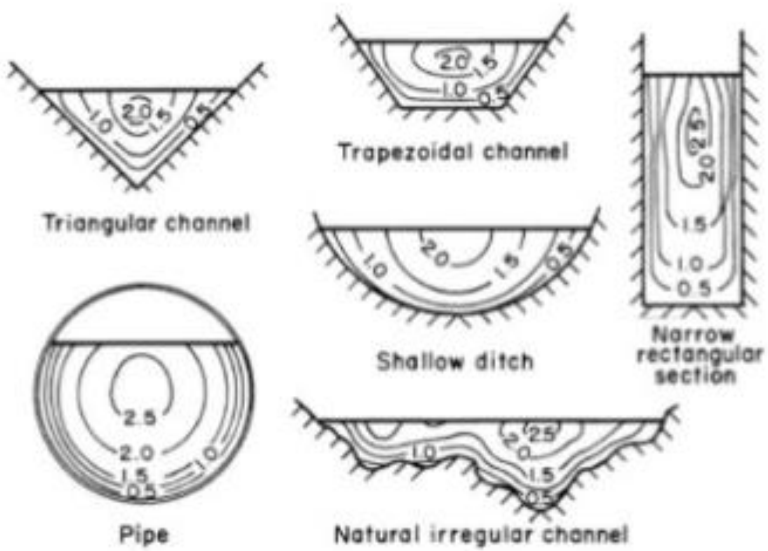
(a) untuk kondisi aliran seragam dan (b) untuk kondisi aliran tidak seragam



Gambar (a) adalah keran air yang dibuka saat awal (bukaan kecil) sehingga air yang mengalir kecepatannya kecil, pada kondisi ini terjadi aliran laminer. Kecepata air meningkat pada Gambar (b) dan Gambar (c) sehingga aliran air berubah menjadi turbulen.

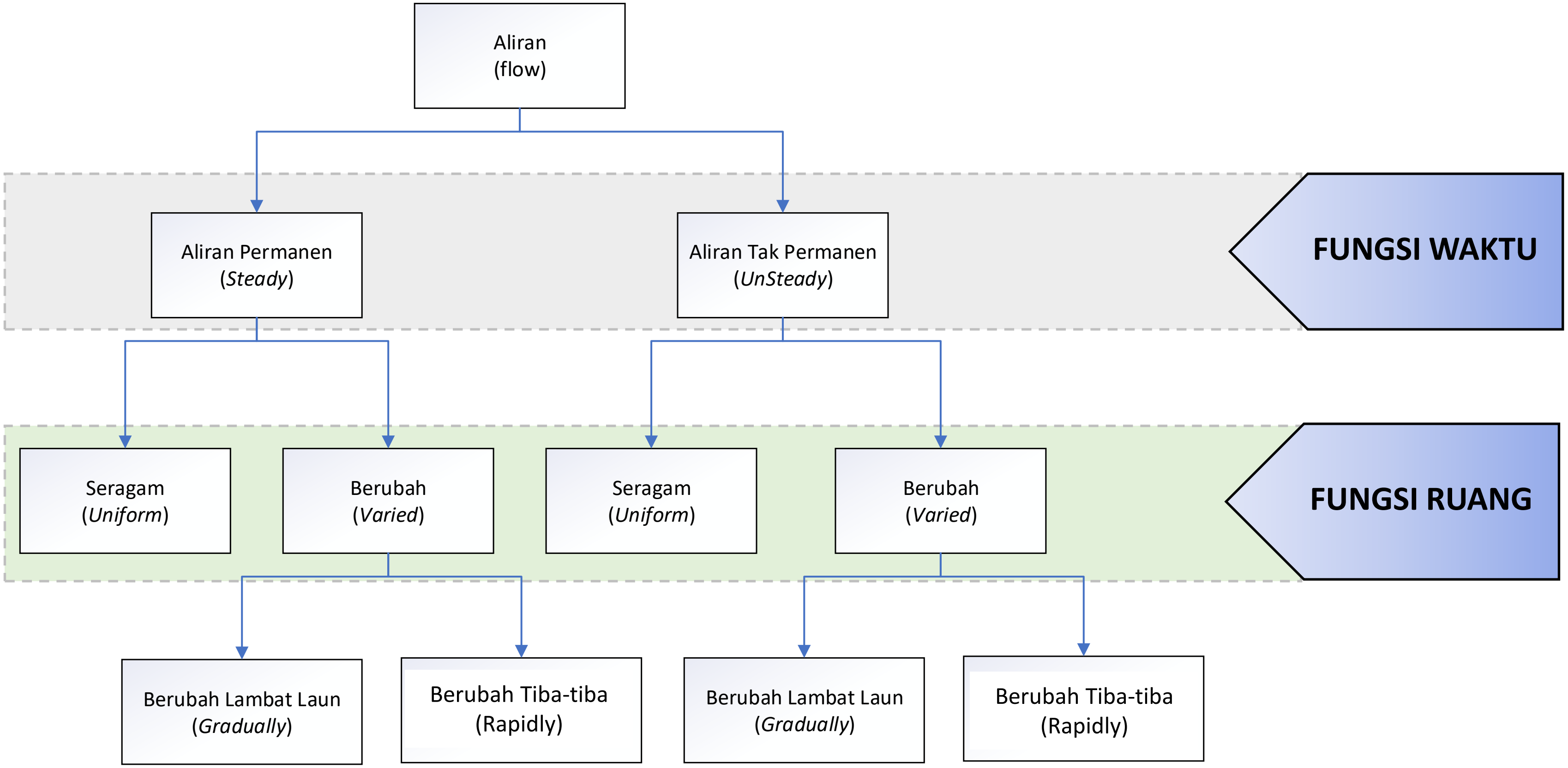


Saluran Terbuka dan Saluran Tertutup



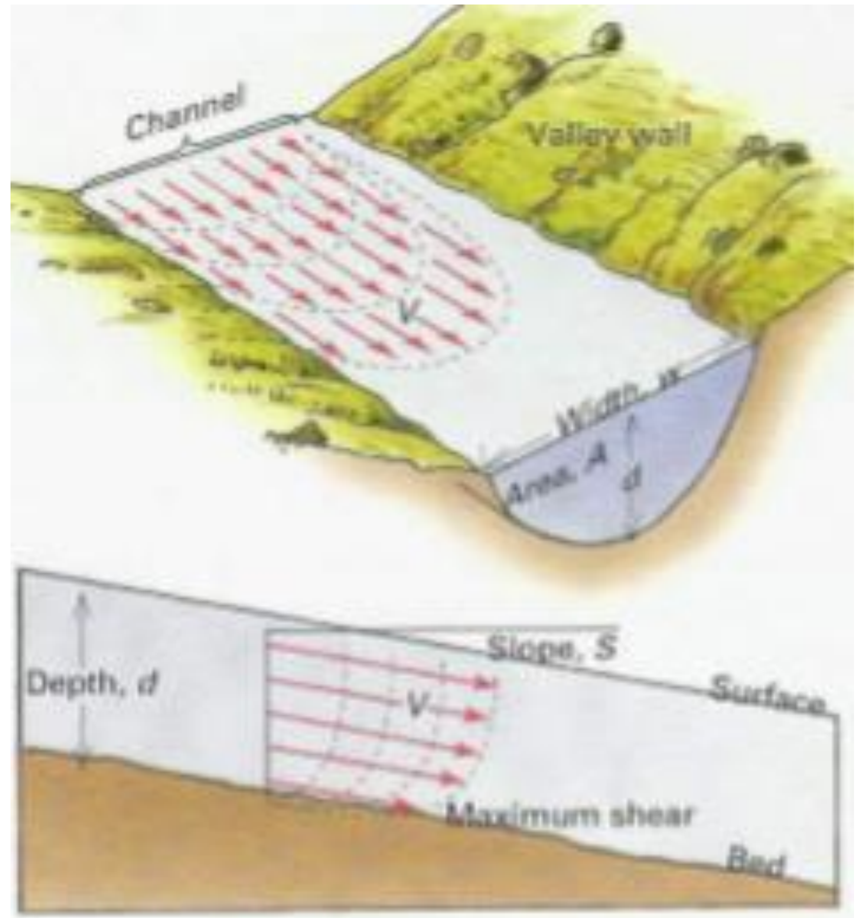
Bentuk Bentuk Potongan Melintang Pada Saluran Terbuka

KLASIFIKASI ALIRAN SALURAN TERBUKA



PENGUKURAN DEBIT
Debit Aliran

DEBIT ALIRAN = jumlah aliran (volume) yang mengalir melalui suatu penampang dalam waktu tertentu, umumnya dinyatakan dalam suatu volume perwaktu(m3/dt).



Q = V x A

Q : debit (m3/dt)

V : Kecepatan Aliran (m/dt)

A : Luas Penampang Aliran (m2)

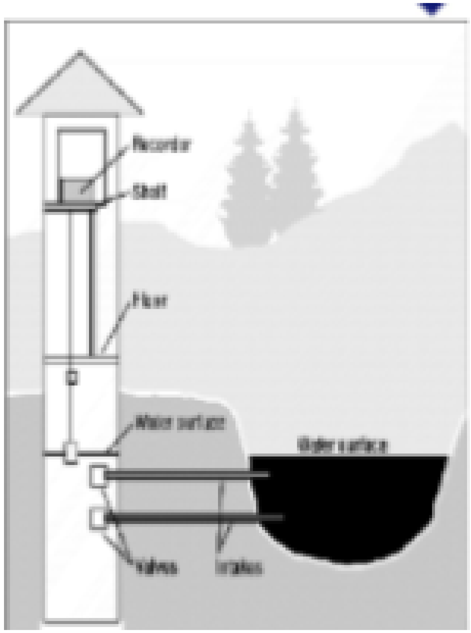
Cara Pengukuran Langsung

1. Volumetrik

Cara volumetrik merupakan cara yang paling sederhana, khususnya pada aliran kecil. Aliran dimasukkan pada bejana kecil atau bejana ukur kemudian dicatat waktunya untuk memenuhi tersebut, sehingga diperoleh debit (V/T).

2. Cara Ambang Ukur

Cara ambang ukur digunakan untuk bangunan air yang mempunyai hubungan tertentu antara debit dengan tinggi muka air. Oleh karena itu, maka setiap bangunan air mempunyai rumus hubungan tertentu tergantung dari lebar (B), tinggi muka air (h) dan tetapan bentuk (n) maupun tetapan debitnya (K). Persamaan umum yang digunakan adalah: **Q = K x B x hn**



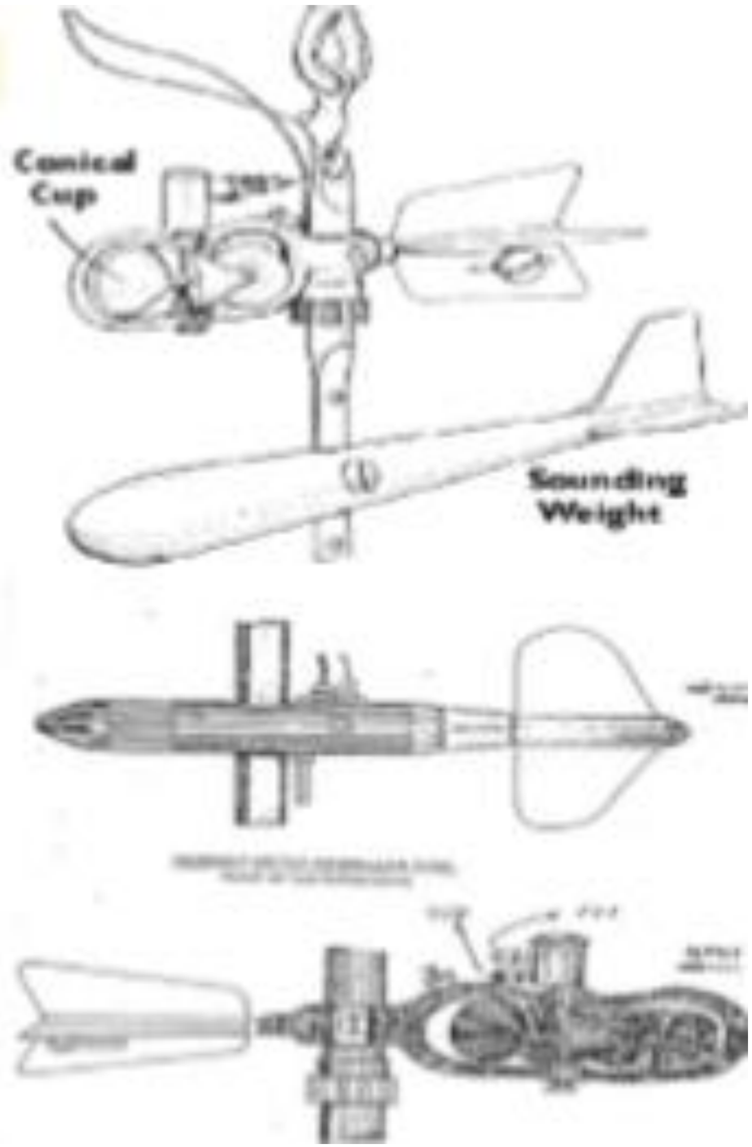
Current meter adalah alat untuk mengukur kecepatan aliran, setiap current meter mempunyai rumus kecepatan aliran,

$$V_{air} = a + Bn$$

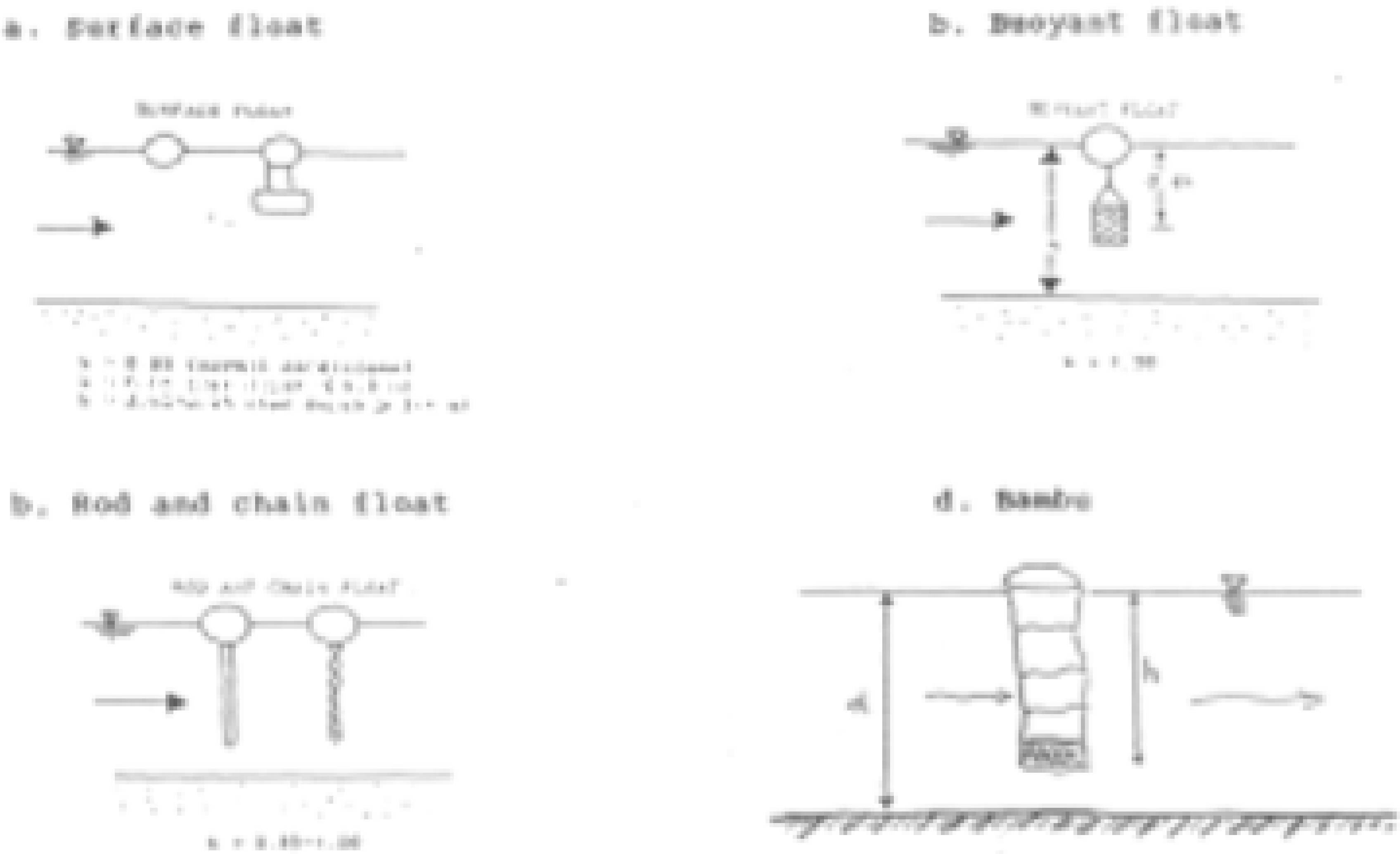
adapun :

a dan b adalah koefisien regresi (setiap current meter besarannya berbeda)

N adalah jumlah putran baling dibagi dengan waktu putaran.

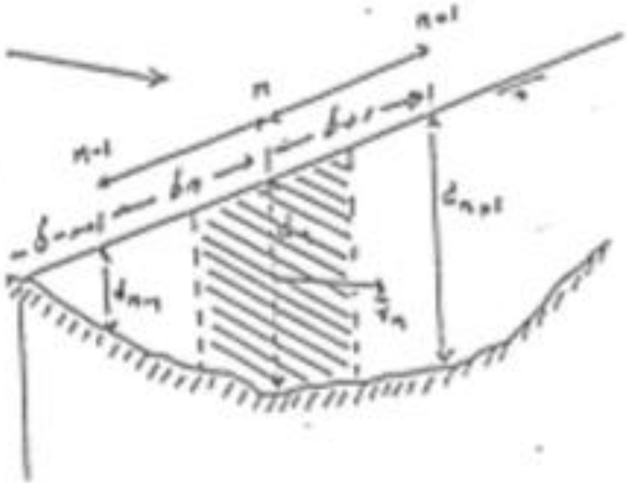


Metode Apung

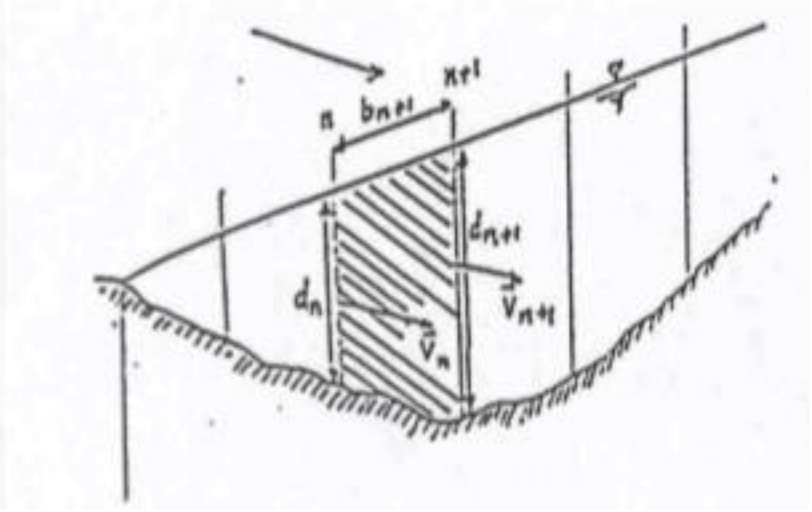


SLOPE-AREA METHOD

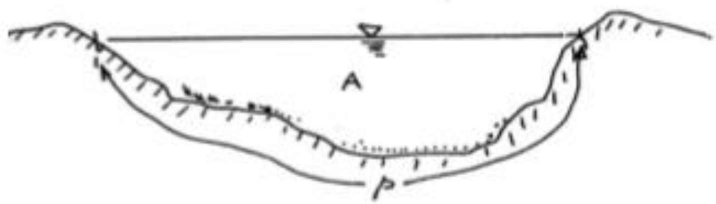
Cara Mid Section

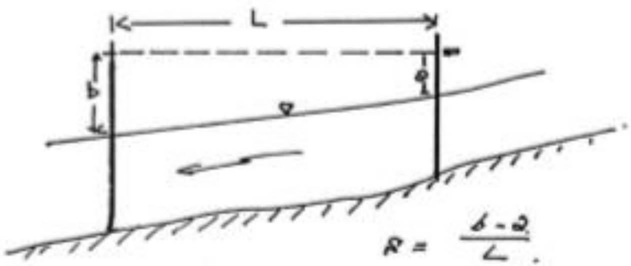


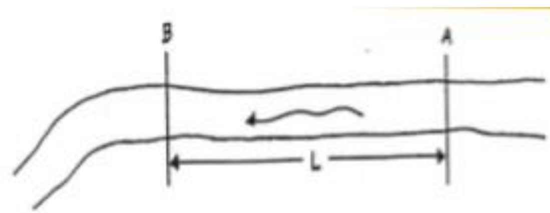
Cara Mean Section



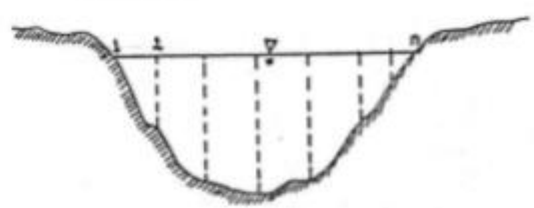
SLOPE-AREA METHOD







Seksi
Pengukuran



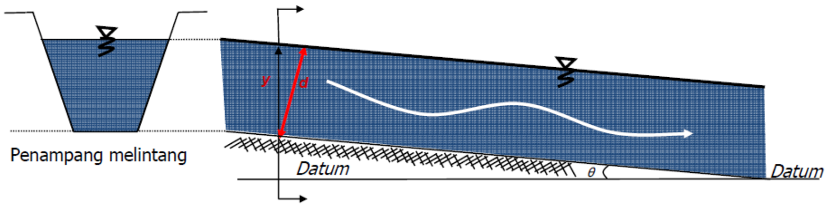
Penampang
Melintang

ELEMEN GEOMETRI

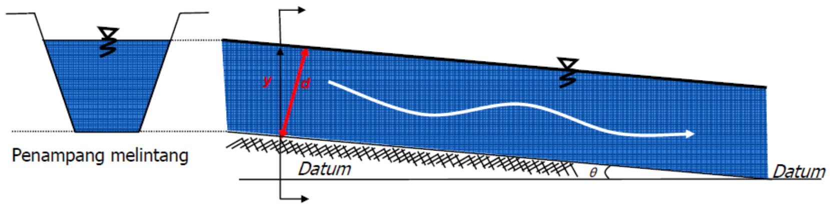
Yang dimaksud dengan **penampang saluran (*channel cross section*)** adalah penampang yang diambil tegak lurus arah aliran, sedang penampang yang diambil vertical disebut **penampang vertical (*vertical section*).**

Dengan demikian apabila dasar saluran terletak horizontal maka penampang saluran akan sama dengan penampang vertikal. Saluran buatan biasanya direncanakan dengan penampang beraturan menurut bentuk geometri yang biasa digunakan

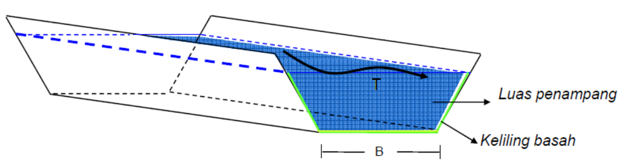
Elemen Geometri penampang memanjang saluran terbuka dapat dilihat sebagai berikut :



Gambar : Penampang Memanjang dan Penampang Melintang aliran saluran terbuka



Gambar 1 : Penampang Melintang dan Penampang Memanjang pada aliran saluran terbuka



Gambar 2 : Parameter Lebar Permukaan (T), Lebar Dasar (B), Luas Penampang dan Keliling Basah Suatu Aliran

dengan notasi ***d*** adalah kedalaman dari penampang aliran, sedang kedalaman **y** adalah kedalaman vertikal (lihat Gambar 1), dalam hal sudut kemiringan dasar saluran sama dengan θ

maka :

$$d = y\cos\theta$$

$$atau$$

$$y = \frac{d}{\cos\theta}$$

Lebar Permukaan adalah lebar penampang saluran pada permukaan bebas (lihat Gb.2). Notasi atau simbol yang digunakan untuk lebar permukaan adalah **T,** dan satuannya adalah satuan panjang.

Notasi atau simbol yang digunakan untuk keliling basah ini adalah **P,** dan satuannya adalah satuan panjang.

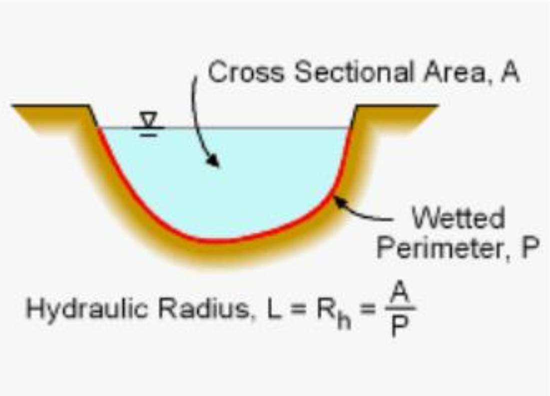
JARI-JARI HIDROLIK

- Jari-jari Hidrolik (*Hidraulic Radius*) dari suatu penampang aliran bukan merupakan karakteristik yang dapat diukur langsung, tetapi sering sekali digunakan didalam perhitungan. Definisi dari jari-jari hidrolik adalah luas penampang dibagi keliling basah, dan oleh karena itu mempunyai satuan panjang; notasi atau simbul yang digunakan adalah **R**, dan satuannya adalah satuan panjang.
- Untuk kondisi aliran yang spesifik, jari-jari hidrolik sering kali dapat dihubungkan langsung dengan parameter geometrik dari saluran. Misalnya, jari-jari hidrolik dari suatu aliran penuh di dalam pipa (penampang lingkaran dengan diameter (D) dapat dihitung besarnya jari-jari hydraulik sebagai berikut:

$$R = \frac{A}{P_w}$$

$$R_{lingkaran} = \frac{\pi . D^2 / 4}{\pi . D} = \frac{D}{4}$$

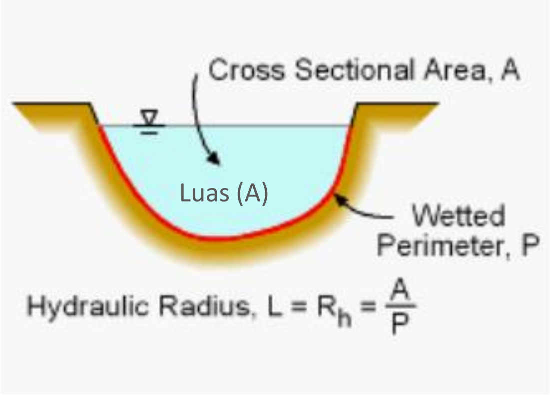
R = Jari-jari hidrolik (ft/m)
A = Luas penampang (ft2 atau m2)
Pw = Keliling basah (ft atau m)
D = Diameter pipa (ft atau m)

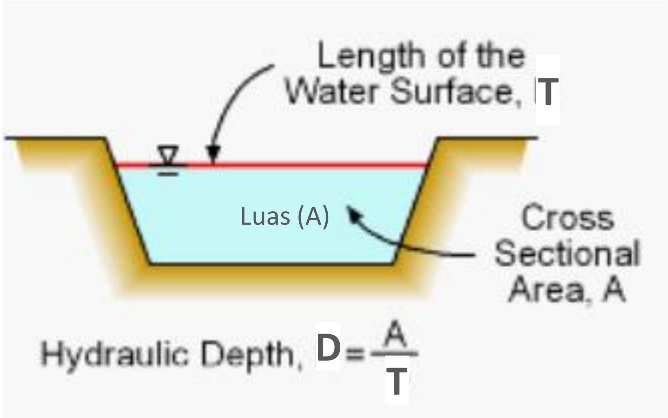


KEDALAMAN HIDROLIK

Kedalaman Hidrolik (***Hydraulic Depth***) dari suatu penampang aliran adalah luas penampang dibagi lebar permukaan, dan oleh karena itu mempunyai satuan panjang. Simbul atau notasi yang digunakan adalah **D**.

$$D = \frac{A}{T}$$





FAKTOR PENAMPANG ALIRAN KRITIS

Faktor Penampang untuk menghitung Aliran Kritis adalah perkalian dari luas penampang aliran A dan akar dari kedalaman hidrolik D. Simbol atau notasi yang digunakan
adalah **Z**.

$$Z = A\sqrt{D}$$

$$= A\sqrt{\frac{A}{T}}$$

FAKTOR PENAMPANG ALIRAN SERAGAM

Faktor penampang untuk perhitungan aliran seragam (Section factor for uniform –flow computation) adalah perkalian dari luas penampang aliran A dan pangkat 2/3 dari jari-jari hydraulik :

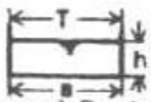
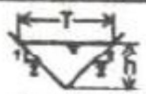
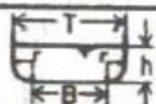
$$\mathbf{A x R^{2/3}}$$

Persamaan / rumus elemen geometri dari berbagai bentuk penampang aliran dapat dilihat pada table berikut :

Penampang	Luas A	Keliling basah O	Jari-jari hidrolis R	Lebar puncak T	Kedalaman hidrolis D	Faktor penampang Z
Persegi Panjang	Bh	$B+2h$	$\frac{Bh}{B+2h}$	B	h	$Bh^{1,5}$
Trapesium	$(B+zh)h$	$B+2h\sqrt{1+z^2}$	$\frac{(B+zh)h}{B+2h\sqrt{1+z^2}}$	$B+2zh$	$\frac{(B+zh)h}{B+2zh}$	$\frac{[(B+zh)h]^{1,5}}{\sqrt{B+2zh}}$
Segi tiga	zh^2	$2h\sqrt{1+z^2}$	$\frac{zh}{2\sqrt{1+z^2}}$	$2zh$	$\frac{1}{2}h$	$\frac{\sqrt{2}}{2}zh^{2,5}$
Lingkaran	$\frac{1}{2}(\theta-\sin\theta)d_0^2$	$\frac{1}{2}\theta d_0$	$\frac{1}{4}\left(1-\frac{\sin\theta}{\theta}\right)d_0$	$(\sin\frac{1}{2}\theta)d_0$ or $2\sqrt{h(d_0-h)}$	$\frac{1}{8}\left(\frac{\theta-\sin\theta}{\sin\frac{1}{2}\theta}\right)d_0$	$\frac{\sqrt{2}}{32}\frac{(\theta-\sin\theta)^{1,5}}{(\sin\frac{1}{2}\theta)^{0,5}}d_0^{2,5}$
Parabola	$\frac{1}{2}Th$	$T+\frac{8}{3}\frac{h^2}{T}$ *	$\frac{2T^2h}{3T^2+8h^2}$ *	$\frac{3}{2}\frac{A}{h}$	$\frac{2}{3}h$	$\frac{2}{9}\sqrt{6}Th^{1,5}$
Persegi panjang sisi dibulatkan	$(\frac{\pi}{2}-2)r^2+(B+2r)h$	$(\pi-2)r+B+2h$	$\frac{(\frac{\pi}{2}-2)r^2+(B+2r)h}{(\pi-2)r+B+2h}$	$B+2r$	$\frac{(\frac{\pi}{2}-2)r^2}{B+2r}+h$	$\frac{[(\frac{\pi}{2}-2)r^2+(B+2r)h]^{1,5}}{\sqrt{B+2r}}$
Segi tiga, dasar dibulatkan	$\frac{T^2}{z4}-\frac{r^2}{z}(1-z\cot^{-1}z)$	$\frac{T}{z}\sqrt{1+z^2}-\frac{2r}{z}(1-z\cot^{-1}z)$	$\frac{A}{O}$	$2[z(h-r)+r\sqrt{1+z^2}]$	$\frac{A}{T}$	$A\sqrt{\frac{A}{T}}$

*) Perkiraan yang paling cocok untuk interval $0<x\leq1$, bila $x=4h/T$. Bila $x>1$, dipakai hubungan $P=(T/2)[\sqrt{1+x^2}+1/x\ \ln(x+\sqrt{1+x^2})]$

UNSUR UNSUR GEOMETRIS PENAMPANG

Penampang	Luas A	Keliling basah O	Jari-jari hidrolis R	Lebar puncak T	Kedalaman hidrolis D	Faktor penampang Z
 Persegi Panjang	Bh	$B+2h$	$\frac{Bh}{B+2h}$	B	h	$Bh^{1,5}$
 Trapesium	$(B+zh)h$	$B+2h\sqrt{1+z^2}$	$\frac{(B+zh)h}{B+2h\sqrt{1+z^2}}$	$B+2zh$	$\frac{(B+zh)h}{B+2zh}$	$\frac{[(B+zh)h]^{1,5}}{\sqrt{B+2zh}}$
 Segi tiga	zh^2	$2h\sqrt{1+z^2}$	$\frac{zh}{2\sqrt{1+z^2}}$	$2zh$	$\frac{1}{2}h$	$\frac{\sqrt{2}}{2}zh^{2,5}$
 Lingkaran	$\frac{1}{2}(\theta-\sin\theta)d_0^2$	$\frac{1}{2}\theta d_0$	$\frac{1}{4}\left(1-\frac{\sin\theta}{\theta}\right)d_0$	$(\sin\frac{1}{2}\theta)d_0$ or $2\sqrt{h(d_0-h)}$	$\frac{1}{8}\left(\frac{\theta-\sin\theta}{\sin\frac{1}{2}\theta}\right)d_0$	$\frac{\sqrt{2}}{32}\frac{(\theta-\sin\theta)^{1,5}}{(\sin\frac{1}{2}\theta)^{0,5}}d_0^{2,5}$
 Parabola	$\frac{1}{2}Th$	$T+\frac{8}{3}\frac{h^2}{T}$ *	$\frac{2T^2h}{3T^2+8h^2}$ *	$\frac{3}{2}\frac{A}{h}$	$\frac{2}{3}h$	$\frac{2}{9}\sqrt{6}Th^{1,5}$
 Persegi panjang sisi dibulatkan	$(\frac{\pi}{2}-2)r^2+(B+2r)h$	$(\pi-2)r+B+2h$	$\frac{(\frac{\pi}{2}-2)r^2+(B+2r)h}{(\pi-2)r+B+2h}$	$B+2r$	$\frac{(\frac{\pi}{2}-2)r^2}{B+2r}+h$	$\frac{[(\frac{\pi}{2}-2)r^2+(B+2r)h]^{1,5}}{\sqrt{B+2r}}$
 Segi tiga, dasar dibulatkan	$\frac{T^2}{4z}-\frac{r^2}{z}(1-z\cot^{-1}z)$	$\frac{T}{z}\sqrt{1+z^2}-\frac{2r}{z}(1-z\cot^{-1}z)$	$\frac{A}{O}$	$2[z(h-r)+r\sqrt{1+z^2}]$	$\frac{A}{T}$	$A\sqrt{\frac{A}{T}}$

*) Perkiraan yang paling cocok untuk interval $0<x\leq 1$, bila $x=4h/T$. Bila $x>1$, dipakai hubungan $P=(T/2)[\sqrt{1+x^2}+1/x\ \ln(x+\sqrt{1+x^2})]$

PENAMPANG SALURAN LEBAR

Penampang Saluran Lebar (*Wide Open Channel)* adalah suatu penampang saluran terbuka yang lebar sekali dimana berlaku pendekatan sebagai saluran terbuka berpenampang persegi empat dengan lebar yang jauh lebih besar daripada kedalaman aliran B >> y, dan keliling basah P disamakan dengan lebar saluran B. Dengan demikian maka luas penampang A = B . y; P =B sehingga :

$$R = \frac{A}{P} = \frac{B_y}{B} = y$$

Debit Aliran (Discharge)

Debit aliran *adalah* volume air yang mengalir melalui suatu penampang tiap satuan waktu, simbol/notasi yang Apabila hukum ketetapan massa digunakan adalah **Q**.

Dalam praktek faktor penting dalam studi hidraulika adalah kecepatan aliran **V** atau debit aliran **Q**

$$\mathbf{Q = A.V}$$

Dengan A adalah luas penampang Aliran

Dengan demikian besarnya debit aliran adalah sudah tertentu. Artinya untuk bisa menghitung debit aliran Q, terlebih dahulu harus dihitung kecepatan V. Rumus kecepatan ini bisa diperoleh secara langsung atapun secara matematis-empiris yaitu berdasarkan percobaan-percobaan : **CHEZY, MANNING dan STRCKLER**

RUMUS CHEZY

Chezy mencari hubungan bahwa zat cair yang melalui saluran terbuka akan menimbulkan tegangan geser (Tahanan) pada dinding saluran, dan akan diimbangi oleh komponen gaya berat yang bekerja pada zat cair dalam arah aliran. Setelah melalui penurunan rumus maka diperoleh persamaan Chezy :

$$V = C\sqrt{R}.\,I$$

Dimana

V adalah Kecapatan Aliran (m/dt)

R =A/P = jarijari hidraulik (m)

I = Kemiringan dasar saluran

C = Koefisien Chezy, A-Luas Basah

P = Keliling basah nilai koefisien Chezy

Menentukan Nilai C secara Empiris

Rumus Kutter :

$$C = \frac{23 + \frac{0{,}00155}{S} + \frac{1}{N}}{1 + \left(23 + \frac{0{,}00155}{S}\right)\frac{N}{\sqrt{R}}}$$

Rumus Bazin

$$C = \frac{87}{1 + \frac{\gamma B}{\sqrt{R}}}$$

Dimana :

N = Angka kekasaran Manning

S = Kemiringan memanjang Saluran

γB = Koefisien yang tergantung pada kekasaran dinding

Koefisien Kekasaran Bazin

Jenis Dinding Saluran	**γB (Koefisien kekasaran Dinding)**
Dinding Sangat Halus (semen)	0,06
Dinding Halus (papan, Batu, Bata)	0,16
Dinding Batu Pecah	0,46
Dinding Tanah Sangat Teratur	0,85
Saluran Tanah dengan kondisi biasa	1,30
Saluran Tanah dengan dasar batu pecah dan tebing rumput	1,73

Sumber : Hidraulika II, Bambang Triatmojo

RUMUS MANNING

$$V = 1/n \; R^{2/3} \; I^{1/2}$$

Dengan *n* adalah koefisien Manning dan *R* adalah jari-jari Hydraulik, yaitu perbandingan antara luas tampang aliran *A* dan keliling basah *P*.
Untuk pipa lingkaran, $A = \pi D^2/4$ dan $P = \pi D$, sehingga:
Atau
$D = 4R$

KOEFISIEN KEKASARAN MANNING

Bahan Dinding Saluran	$\boldsymbol{Koefisien\ Manning\ (n)}$
Besi Tuang dilapis	0,014
Kaca	0,010
Saluran Beton	0,013
Bata dilapis mortar	0,015
Pasangan Batu disemen	0,025
Saluran Tanah bersih	0,022
Saluran Tanah	0,030
Saluran dengan dasar batu dan Tebing Rumput	0,040
Saluran pada galian batu padas	0,040

Sumber : Hidraulika II, Bambang Triatmojo

RUMUS STRCKLER

Rumus Strickler yang banyak digunakan pada pengaliran di saluran terbuka, juga berlaku untuk pengaliran di pipa. Rumus tersebut mempunyai bentuk:

$$\mathbf{V = k \ R^{2/3} \ I^{1/2}}$$

Dengan k adalah koefisien Strickler dan *R* adalah jari-jari Hydraulik, yaitu perbandingan antara luas tampang aliran *A* dan keliling basah *P*.

Angka kekasaran STRICKLER

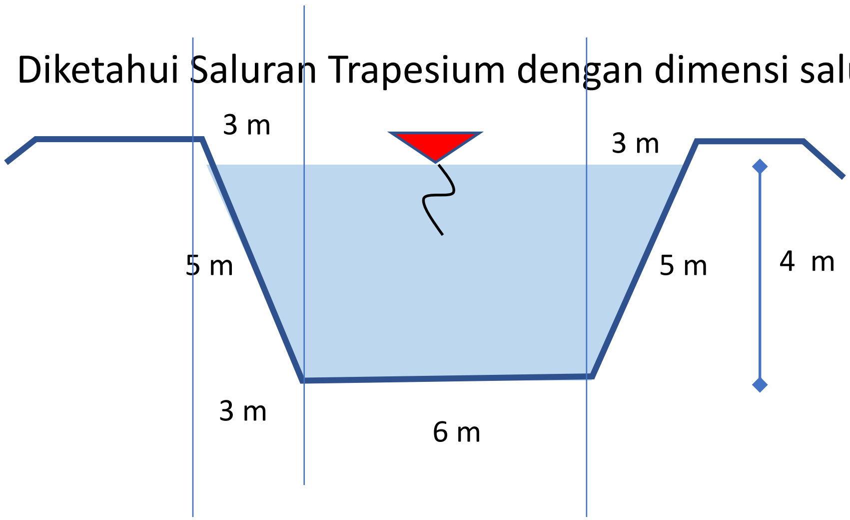
- saluran pasangan 60
- saluran beton 70
- saluran tanah bersih 30-45
- saluran bersemen plesteran 70

Tabel 3.1. Harga – harga kekasaran koefisien Strickler (k) untuk saluran – saluran tanah

Debit rencana m^3/dt	**k** $m^{1/3}/dt$
$Q > 10$	45
$5 < Q < 10$	42,5
$1 < Q < 5$	40
$Q < 1$	35

APLIKASI PERMASAL

Diketahui Saluran Trapesium dengan dimensi saluran se



RUMUS CHEZY

Chezy mencari hubungan bahwa zat cair yang melalui saluran terbuka akan me tegangan geser (Tahanan) pada dinding saluran, dan akan diimbangi oleh komp berat yang bekerja pada zat cair dalam arah aliran. Setelah melalui penurunan ru diperoleh persamaan Chezy :

$$V = C\sqrt{R.I}$$

Dimana

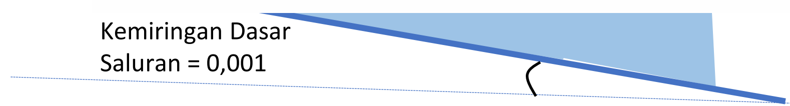
V adalah Kecapatan Aliran (m/dt)

R =A/P = jarijari hidraulik (m)

I = Kemiringan dasar saluran

C = Koefisien Chezy, A-Luas Basah

P = Keliling basah nilai koefisien Chezy



Hitunglah Debit Aliran Jika :

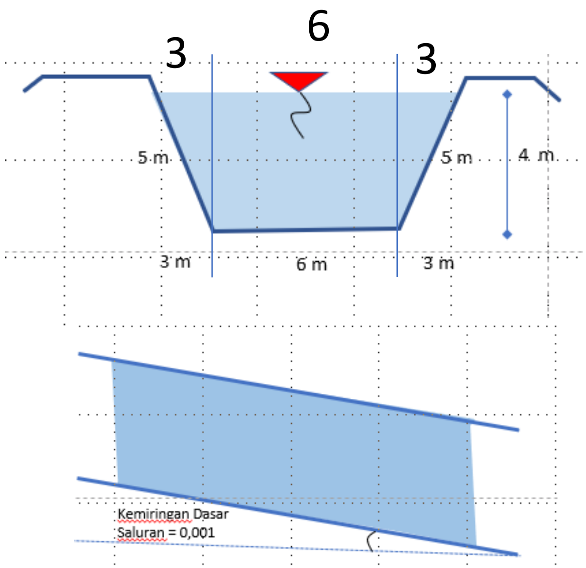
a. Menggunakaan rumus Chezy, dimana Koefisien Chezy = 50
b. Menggunakan Rumus Bazin, dengan Nilai C = 45,80, selain mencari debit carilah koefisien kekasaran dinding $\gamma B\ dan\ tetntukan\ apa\ jenis\ salurannya$
c. Menggunakan Rumus Ganguillet Kutter, jenis dinding saluran pasangan batu disemen.
d. Dengan rumus Manning, Jenis dinding saluran pasangan batu disemen
e. Dengan Rumus Strickler, jenis dinding saluran pasangan batu disemen

JAWAB

$$A = \frac{6+12}{2} \times 4 \; = 36 \text{ m}^2$$

$$P = 5 + 6 + 5 = 16$$

$$\boldsymbol{R} = \frac{A}{P} = \frac{36}{16} \quad = \mathbf{2{,}25\ m}$$



Jawaban :

a. Menggunakaan rumus Chezy, dimana Koefisien Chezy = 50

$$Q = A.V = A.C\sqrt{R.I} = 36 \times 50 \times \sqrt{2{,}25 . 0{,}001} = \mathbf{85{,}381\ m3/dt}$$

b. Menggunakan Rumus Bazin, dengan Nilai C = 45,80, selain mencari debit carilah koefisien kekasaran dinding $\gamma B \; dan \; tetntukan \; apa \; jenis \; salurannya$

C = 45,80

$$C = \frac{87}{1+\frac{\gamma B}{\sqrt{R}}} \rightarrow 1 + \frac{\gamma B}{\sqrt{R}} = \frac{87}{C} \rightarrow \gamma B = \left[\frac{87}{C} - 1\right]\sqrt{R}$$

$$\gamma B = \left[\frac{87}{C} - 1\right]\sqrt{2{,}25} = 1{,}3485$$

Dari Tabel Harga Koefisien Kekasaran Bazin Jenis Saluran yang mendekati dengan nilai γB yang didapat yaitu 1,3485 termasuk dalam jenis **Saluran Tanah dengan Kondisi Biasa.**

$$Q = A.V = A.C\sqrt{R}.I = 36.45{,}80\sqrt{2{,}25 \; x \; 0{,}001} = 78{,}209 \; m3/dt$$

c. Menggunakan Rumus Ganguillet Kutter, jenis **dinding saluran pasangan batu disemen.**
d. Dengan rumus Manning, Jenis dinding saluran pasangan batu disemen
e. Dengan Rumus Strickler, jenis dinding saluran pasangan batu disemen

Jawaban c.

n = 0,025 (sesuai Tabel)

$$C = \frac{23+\frac{0,00155}{I}+\frac{1}{n}}{1+\left[23+\frac{0,00155}{I}\right]\frac{n}{\sqrt{R}}} = \frac{23+\frac{0,00155}{0,001}+\frac{I}{0,025}}{1+\left[23+\frac{0,00155}{0,001}\right]\frac{n}{\sqrt{2,25}}} = 45,807$$

Q = A.V = A.C$\sqrt{R.I} = 36 \times 45,807\sqrt{2,25 \times 0,001}$ = 78,221 m3/dt

Jawaban d.

n = 0,025 (sesuai Tabel)

C = $\frac{I}{n} \times R^{1/6} = {}^{I}/_{0,025} \times 2,25^{1/6}$ = 45,788

Q = A.V = A.C$\sqrt{R.I} = 36 \times 45,788\sqrt{2,25 \times 0,001}$ = 78,189 m3/dt

Jawaban e.

n=0,025

k = ${}^{l}/_{n} = {}^{l}/_{0,025} = 40$

V = $k \times R^{2/3} \times I^{1/2} = 40 \times 2,25^{2/3} \times 0,001^{1/2} = 2,172\ m/dt$

Q = A.V = $36 \times 2,172$ = 78,192 m3/dt

Pustaka

- Lily Montarsih Limantara, Rekayasa Hidrologi
- Imam Subarkah, Hidrologi Untuk Perencanaan Bangunan Air
- Bambang Triatmodjo, Hidarulika
- Ranga Raju, Aliran Melalui Saluran Terbuka
- Reuben MO dan Steven JW, Dasar-dasar Mekanika Fluida Teknik
- Suripin, Mekanika Fluida dan Hidraulika Saluran Terbuka untuk Teknik Sipil

S E L E S A I

MATERI – 1

PENGANTAR HIDROLOGI

Disampaikan Oleh :
IRahardjo Samiono S.T., M.T
Muhamad Komarudin S.Si., M.Si

Materi Pembelajaran

Mata Kuliah Hidrologi PTS-FTSP, SKS : 2
(Ceramah & Diskusi, Tugas, Tutorial, Ujian)

DOSEN :

1. Muhamad Komarudin S.Si., M.Si.
2. Rahardjo Samiono S.T., M.T.

Materi Pertemuan

No	Materi
1	Pengantar MK Hidrologi
2	Hujan (Presipitasi)
3	Evaporasi
4	Infiltrasi
5	Air Tanah
6	Limpasan
7	Sungai dan Danau
8	UTS

No	Materi
9	Perhitungan Debit Aliran
10	Analisa Kualitas Air
11	Perhitungan Distribusi Hujan
12	Metode Rasional
13	Metode Rasional
14	Neraca Air
15	Perencanaan Saluran
16	UAS

Bobot Penilaian

1	**Ujian Tengah Semester**	**30 %**
2	Ujian Akhir Semester	30 %
3	Tugas Individu	15 %
4	Tugas Kelompok	15 %
5	Kehadiran	10 %
	Total	100 %

Pustaka

- Linsley, R.K., Kohler, M.A., Paulhus, J.L., 1975. Hydrology for Engineers. 2nd. Ed Mc Graw Hill Kogakusha Ltd. Tokyo
- Seyhan, E., 1977 Fundamental Hydrology. Geografisch Institute der Rijks Universitiet Utrecr. The Netherlands
- Todd, D.K. 2005. Groundwater Hydrology. John Willey & Sons Inc
- Tim Davie, Fundamentals of Hydrology
- Chay Asdak, Hidrologi dan Pengelolaan Daerah Aliran Sungai

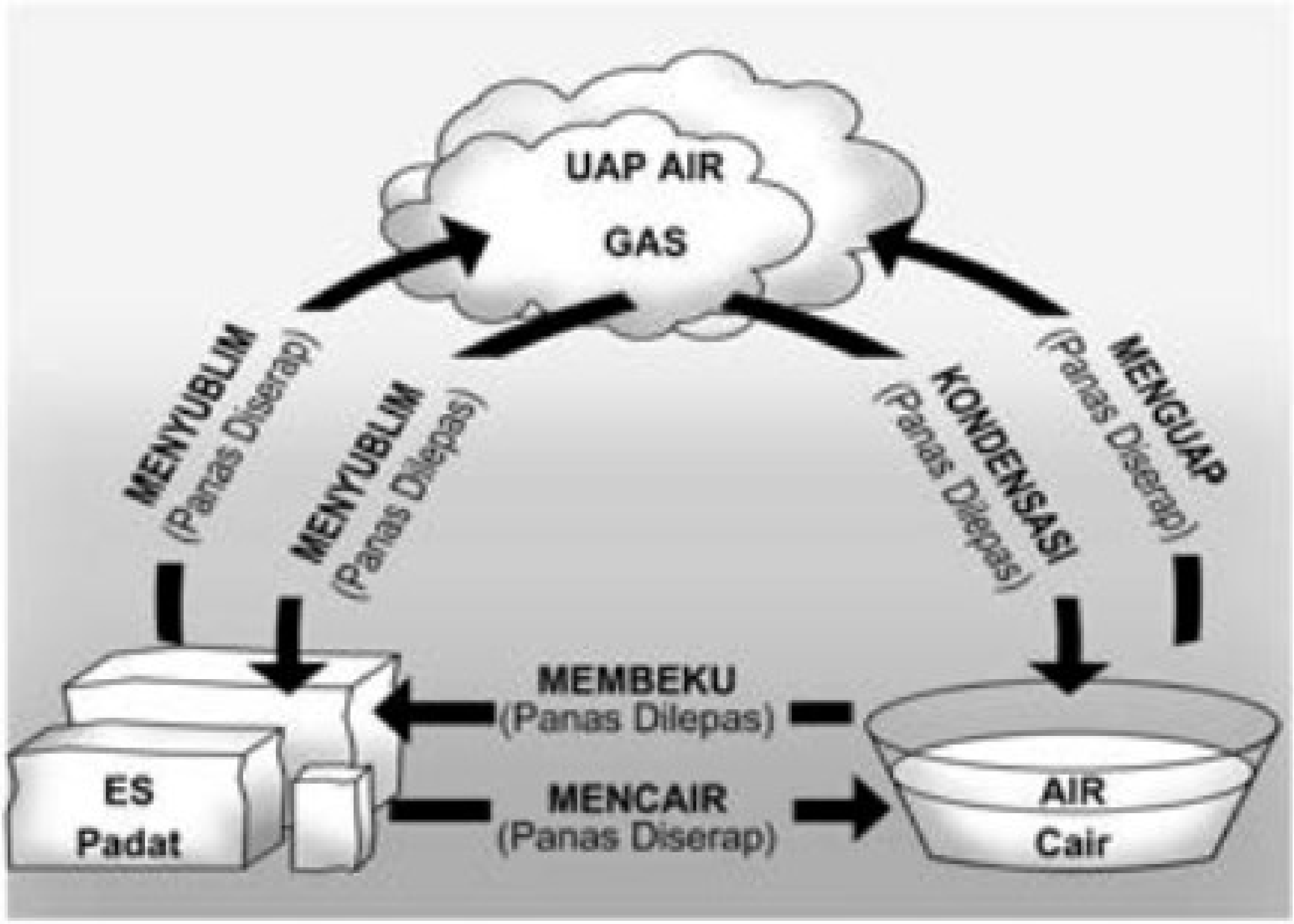
ISTN

MATERI – 1

SIKLUS HIDROLOGI

Disampaikan Oleh :
Ir. Rahardjo Samiono M.T
Muhamad Komarudin S.Si., M.Si

Bentuk Air



UAP AIR
GAS
MENYUBLIM
(Panas Diserap)
MENYUBLIM
(Panas Dilepas)
KONDENSASI
(Panas Dilepas)
MENGUAP
(Panas Diserap)
MEMBEKU
(Panas Dilepas)
MENCAIR
(Panas Diserap)
ES
Padat
AIR
Cair

Pengertian Hidrologi

- Hidrologi merupakan ilmu geografi fisik yang berkaitan dengan keberadaan air di muka bumi sebatas pada lapisan kehidupan dengan sorotan khusus pada sifat, fenomena dan distribusi air di daratan
- Hidrologi dikategorikan secara khusus mempelajari kejadian air di daratan/bumi, deskripsi pengaruh sifat daratan terhadap air, pengaruh fisik air terhadap daratan dan mempelajari hubungan air dengan kehidupan

Ruang Lingkup Pembelajaran Hidrologi

- Pengukuran, pencatanan dan publikasi data air
- Deskripsi tentang sifat, fenomena dan distibusi air menurut ruang maupun waktu
- Analisis data air untuk membangun teori-teori yang ada dalam hidrologi
- Aplikasi teori-teori hidrologi untuk memecahkan masalah-masalah praktis yang berkaitan dengan air seperti banjir, kekeringan, pengelolaan DAS dan Penyediaan Air

Secara etimologi hidrologi berasal dari bahasa Yunani, Hydrologia, yang artinya ilmu air. Sehingga hidrologi merupakan cabang dari ilmu geografi yang mengkaji mengenai pergerakan, distribusi dan juga kualitas air di bumi, kajian ini juga meliputi siklus hidrologi dan sumber daya air. Lebih dalam lagi ilmu hidrologi mengkaji tentang :

1. hidrometeorologi (air yang ada di udara dengan wujud gas),
2. potamologi (aliran permukaan),
3. kriologi (air dengan wujud padat contohnya es dan salju),
4. geohidrologi (air tanah), serta
5. limnologi (air permukaan yang cenderung tenang contohnya danau, dan waduk).

Perkiraan Jumlah Air di Bumi

	Volume (×10³ km³)	Percentage of total
Oceans and seas	1,338,000	96.54
Ice caps and glaciers	24,064	1.74
Groundwater	23,400	1.69
Permafrost	300	0.022
Lakes	176	0.013
Soil	16.5	0.001
Atmosphere	12.9	0.0009
Marsh/wetlands	11.5	0.0008
Rivers	2.12	0.00015
Biota	1.12	0.00008
Total	1,385,984	100.00

Source: Data from Shiklomanov and Sokolov (1983)

Perkiraan Jumlah Air di Indonesia

Menurut data Ditjen Sumberdaya Alam Kementerian PU (2003), Ketersediaan air menurut distribusi pulau terbesar ada di Pulau Kalimantan dan terkecil di Nusa Tenggara hal ini dipengaruhi oleh kondisi ekoregion di masing-masing pulau. Ketersediaan Air menurut Pulau Di Indonesia

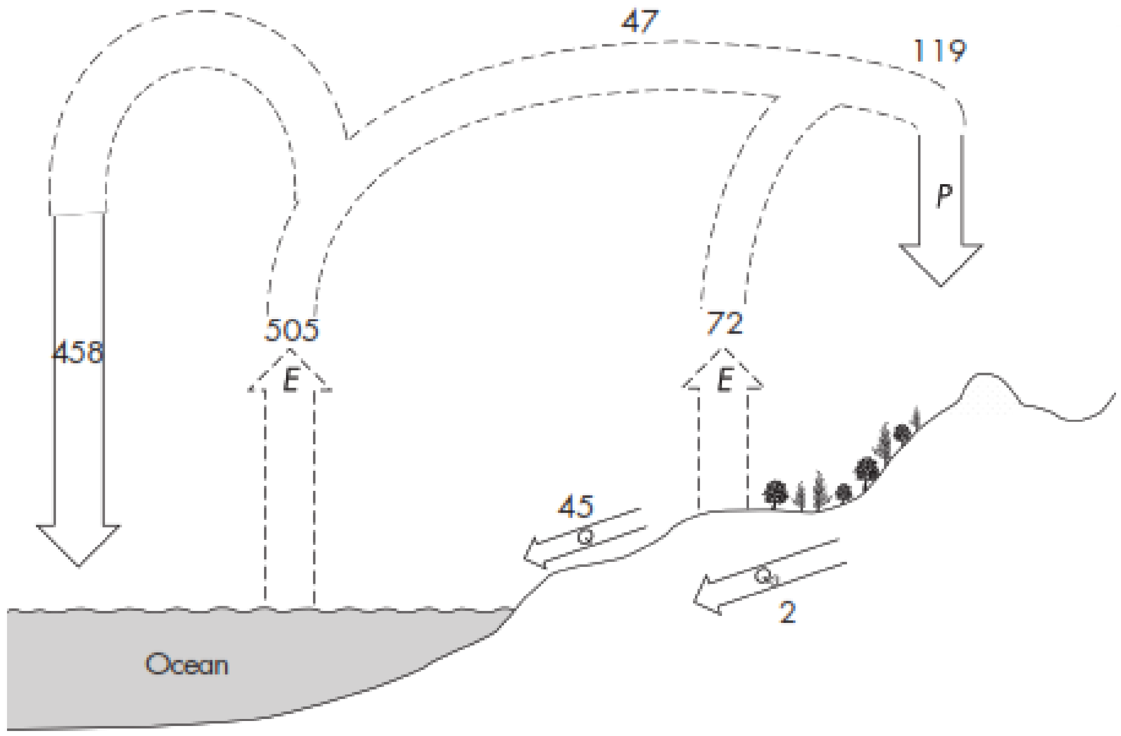
No	Pulau	Ketersediaan Air (juta m3/tahun)			
		Musim Hujan	Musim Kemarau	Total	% Total Nasional
1	Sumatera	384.774,4	96.193,6	480.968,0	25%
2	Jawa & Bali	101.160,8	25.290,2	126.451,0	7%
3	Kalimantan	389.689,3	167.009,7	556.699,0	28%
4	Sulawesi	129.400	14.377,8	143.778,0	7%
5	Nusa Tenggara	37.940,4	4.215,6	42.156,0	2%
6	Papua	381.763,9	163.613,1	545.377,0	28%

Sumber : Ditjen Sumber Daya Alam Kementerian PU (2003)

SIKLUS HIDROLOGI

- Sebagai titik awal untuk mempelajari hidrologi, siklus hidrologi merupakan konsep penting untuk dipahami.
- Siklus Hidrologi merupakan model konseptual tentang bagaimana air bergerak di antara bumi dan atmosfer dalam berbagai keadaan sebagai gas, cair, atau padat.
- Seperti halnya model konseptual apa pun, model ini mengandung banyak penyederhanaan yang bersifat umum.
- Ada berbagai skala yang dapat dilihat dari siklus hidrologi, tetapi sangat membantu untuk memulai dari skala global yang besar dan kemudian pindah ke unit hidrologi yang lebih kecil dari cekungan sungai atau daerah tangkapan air.

SIKLUS HIDROLOGI



The global hydrological cycle. The numbers represent estimates on the total amount of water (thousands of km3) in each process per annum. E = evaporation; P = precipitation; QG = subsurface runoff; Q = surface runoff.

Source: Redrawn from Shiklomanov (1993)

Processes in the hydrological cycle operating at the basin or catchment scale. Q = runoff; the subscript G stands for groundwater flow; TF for throughflow; I = interception; E = evaporation; P = precipitation.

WATER BALANCE

Persamaan Umum Neraca Air (Water Balance)

$$P \pm E \pm \Delta S \pm Q = 0$$

Dimana :

P = Hujan / Presipitasi

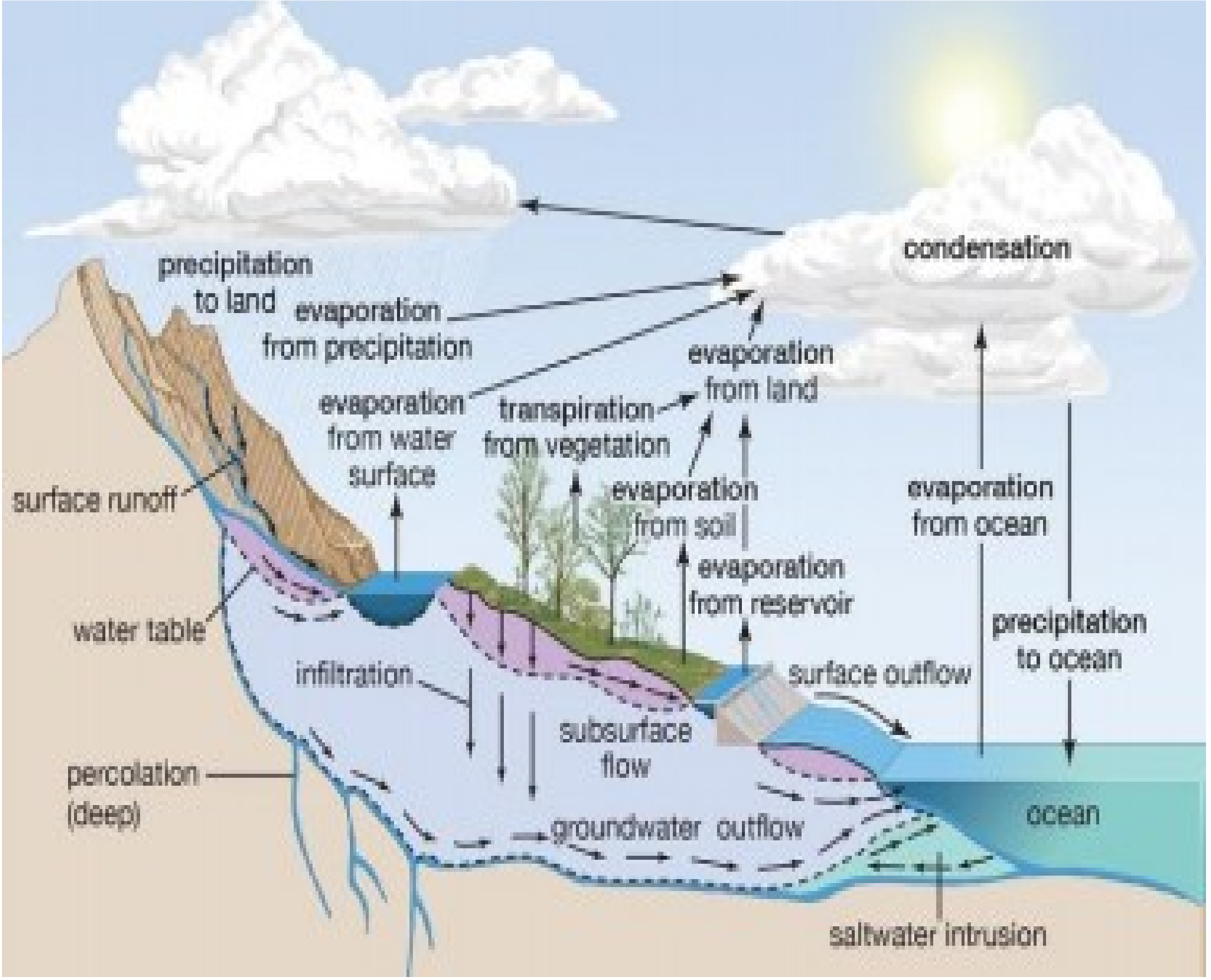
E = Evaporasi

ΔS = Perubahan Simpanan Air

Q = Limpasan Permukaan

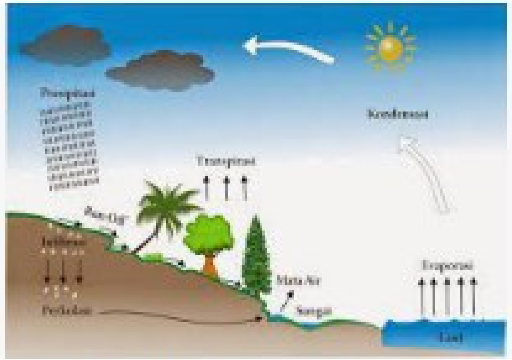
SIKLUS HIDROLOGI

Sirkulasi air yang berpola siklus itu tidak pernah berhenti dari atmosfer ke bumi dan kembali ke atmosfer melalui kondensasi, presipitasi, evaporasi, dan transpirasi. Pemanasan air samudra oleh sinar matahari merupakan kunci proses siklus hidrologi tersebut dapat berjalan secara kontinu. Air berevaporasi, kemudian jatuh sebagai presipitasi dalam bentuk hujan air, hujan es (padatan maupun berupa salju), maupun kabut. Pada waktu pergerakan dari udara saat menjadi uap menuju bumi, air dapat berevaporasi kembali ke udara. Pada kejadian hujan yang jatuh pada lahan bervegetasi maka air tidak langsung jatuh ke permukaan tanah, tetapi jatuh pada vegetasi yang kemudian dinamakan intersepsi, yaitu air melalui pepohonan diintersepsi oleh tanaman sebelum akhirnya mencapai permukaan tanah. Penyebaran air dari suatu tempat ke tempat yang lain di bumi mengikuti suatu siklus hidrologi. Siklus ini terjadi karena adanya beberapa proses hidrologi yaitu Evaporasi/evapotranspirasi, Presipitasi, Infiltrasi, Perkolasi, Aliran Permukaan *Run Off.*

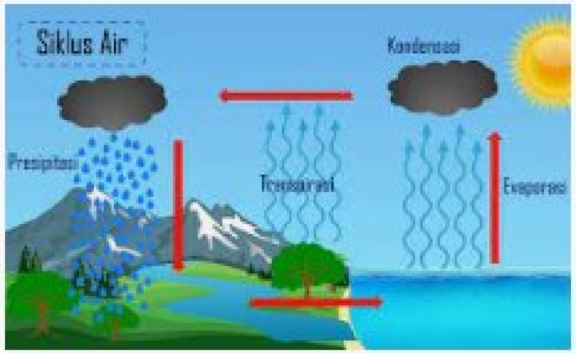


SIKLUS HIDROLOGI

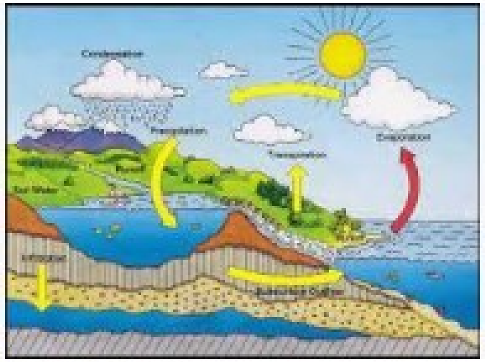
GeoHepi
Siklus Hidrologi - GeoHepi



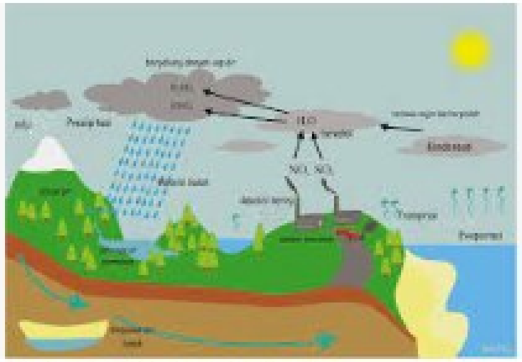
GEOGRAPHY NOTE BOOK - WordPress.com
Siklus Hidrologi – GEOGRAPHY NOTE BOOK



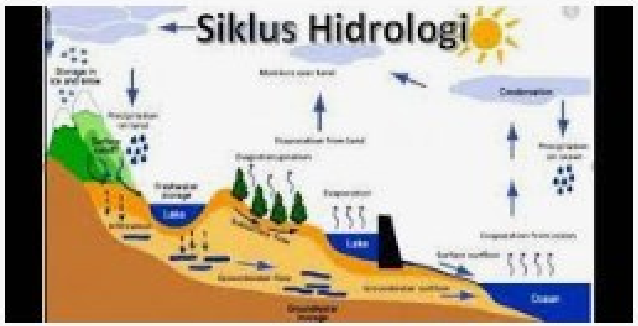
Tokopedia
Siklus Hidrologi Lengkap dengan Gambar ...



IlmuGeografi.com -
Siklus Hidrologi : Pengertian, Tahapan ...



BALAI PSDA PEMALI COMAL
Hidrologi – BALAI PSDA PEMALI COMAL



Ayo Guru Berbagi - Kementerian Pendidikan, Kebudayaan, ...
GURU BERBAGI | Siklus Hidrologi

DISKUSI

- Diskusikan Arti penting Air bagi kehidupan
- Bagaimana ketersediaan air di Indonesia, wilayah mana yang rentan kesediaannya, bagaimana bisa terjadi?
- Uraikan siklus hidrologi menurut anda

MATERI – 2

HUJAN (PRESIPITASI)

Disampaikan Oleh :
Ir. Rahardjo Samiono M.T
Muhamad Komarudin S.Si., M.Si

Pengertian Hujan

Hujan adalah peristiwa turunnya butir-butir air dari langit ke permukaan bumi. Hujan juga merupakan siklus air di bumi.

Curah Hujan

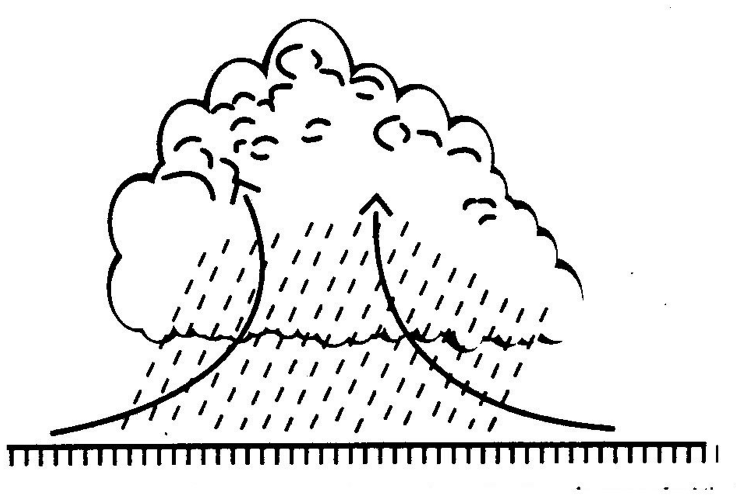
Curah hujan adalah jumlah air yang jatuh di permukaan tanah datar selama periode tertentu yang diukur dengan satuan tinggi (mm).

Tipe Hujan

- Hujan terjadi karena udara basah yang naik ke atmosfer mengalami pendinginan sehingga terjadi proses kondensasi.
- Naiknya udara ke atas dapat terjadi secara siklonik, orografik dan konvektif.

Hujan Konvektif

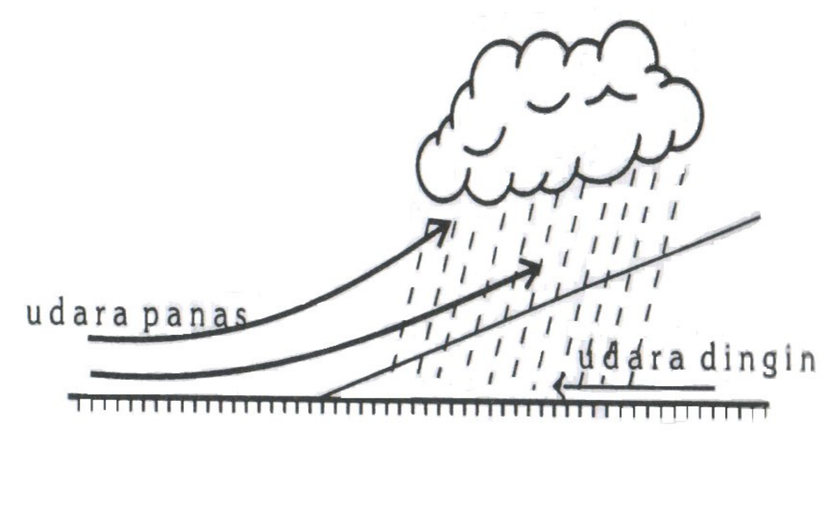
- Hujan jenis ini biasanya terjadi sebagai hujan dengan intensitas yang tinggi, akibat massa udara yang terangkat ke atas oleh pemanasan lahan. Hujan jenis ini biasanya terjadi di daerah yang relatif luas dan bergerak sesuai dengan pergerakan angin.



Pembentukan hujan konvektif

Hujan Sinklonik

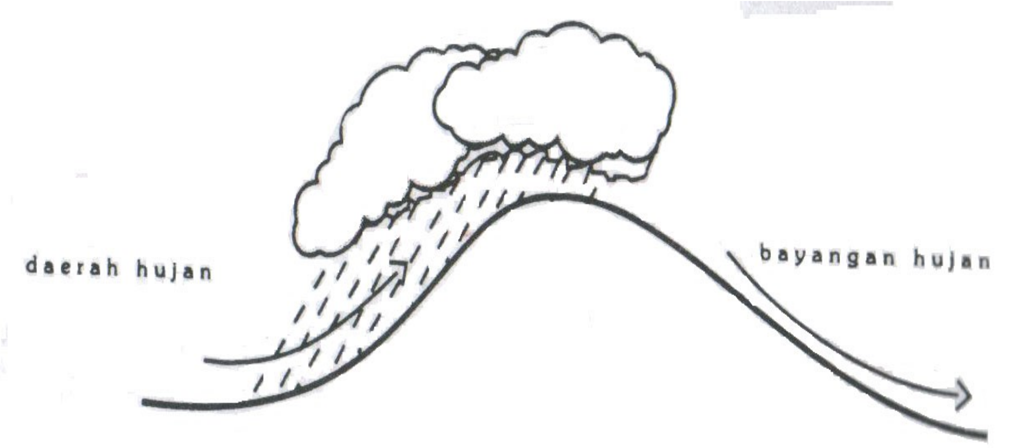
- Hujan jenis ini biasanya terjadi karena udara lembab panas terangkat ke atas oleh lapisan udara yang lebih dingin dan lebih rapat. Penyebaran hujan jenis ini sangat dipengaruhi oleh landai pertemuan antara udara panas dan dingin dan biasanya merupakan hujan dengan daerah penyebaran terbatas dan dalam waktu pendek.



Pembentukan hujan siklonik

Hujan Orografik

- Hujan jenis ini terjadi karena massa udara lembab terangkat ke atas oleh angin karena adanya gunung/pegunungan. Udara lembab yang melintasi daerah pegunungan akan naik dan mengalami pendinginan, sehingga terbentuk awan dan hujan.



Pembentukan hujan orografik

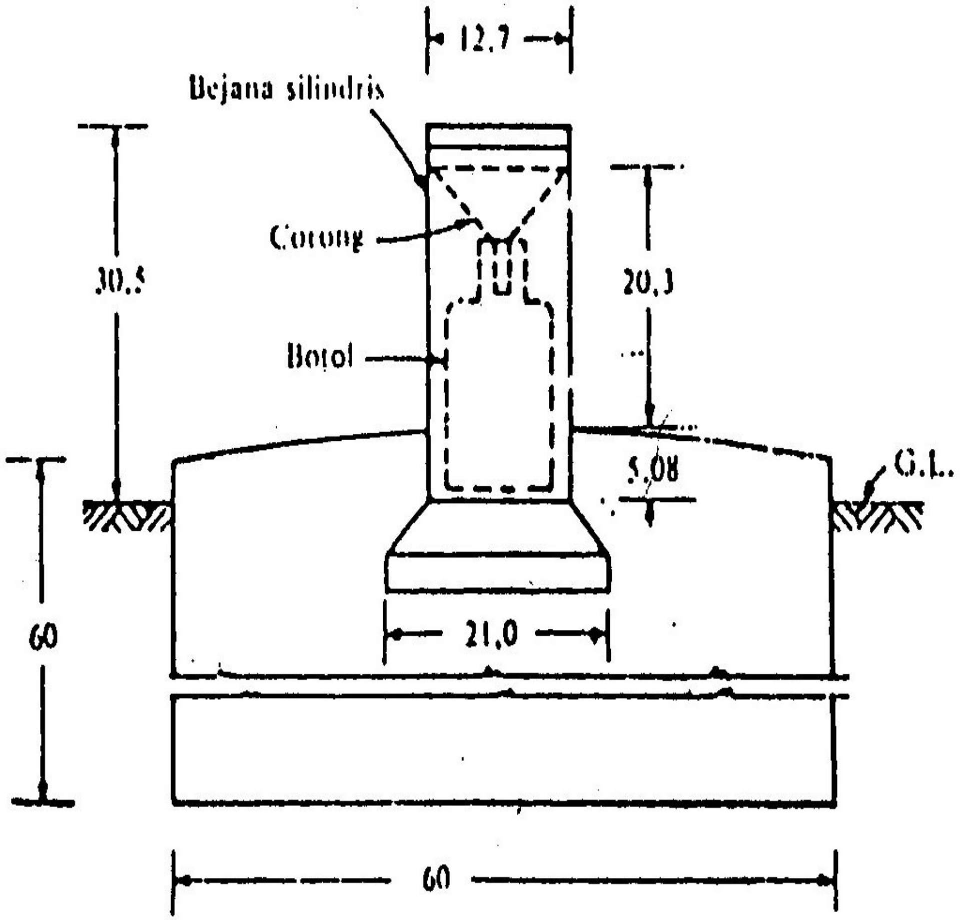
Alat Pengukur Hujan

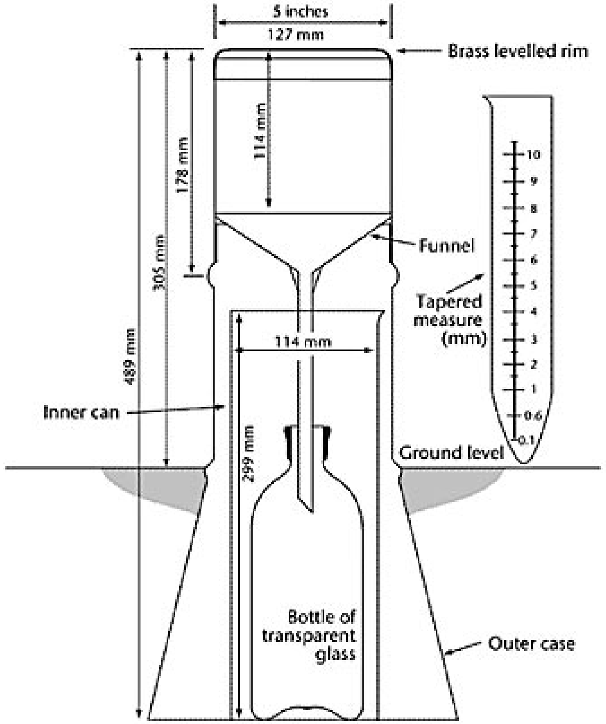
- Alat ukur hujan dapat dibedakan menjadi 2 macam, yaitu penakar hujan biasa (*manual raingauge*) dan penakar hujan otomatis (*automatic raingauge*).
- Data curah hujan dapat berupa data curah hujan harian atau curah hujan pada periode waktu yang lebih pendek, misal setiap menit. Data hujan tipe pertama dapat diukur dengan penakar hujan biasa terdiri dari bejana dan corong seluas 200 cm^2 yang dipasang setinggi 120 cm dari permukaan tanah. Data hujan untuk periode pendek didapat dari alat penakar hujan otomatis ARR (*automatic rainfall recorder*) yang dapat merekam setiap kejadian hujan selama jangka waktu tertentu. Berdasarkan mekanisme perekaman data hujan ada tiga jenis ARR, yaitu tipe *weighing bucket*, *tipping bucket* dan *float*.

Stasiun Hujan

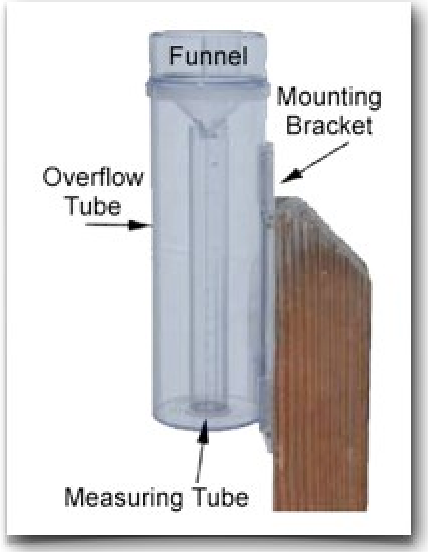
Alat Penakar Hujan Biasa

Alat penakar hujan biasa terdiri dari corong dan botol penampung yang berada di dalam suatu tabung silinder. Hujan yang jatuh pada corong akan tertampung di dalam tabung silinder, kemudian kedalaman hujan di dapat dari pengukuran volume air yang tertampung dan luas corongnya. Curah hujan kurang dari 0,1 mm dicatat sebagai 0,0 mm, sedangkan jika tidak ada hujan dicatat dengan garis (-).





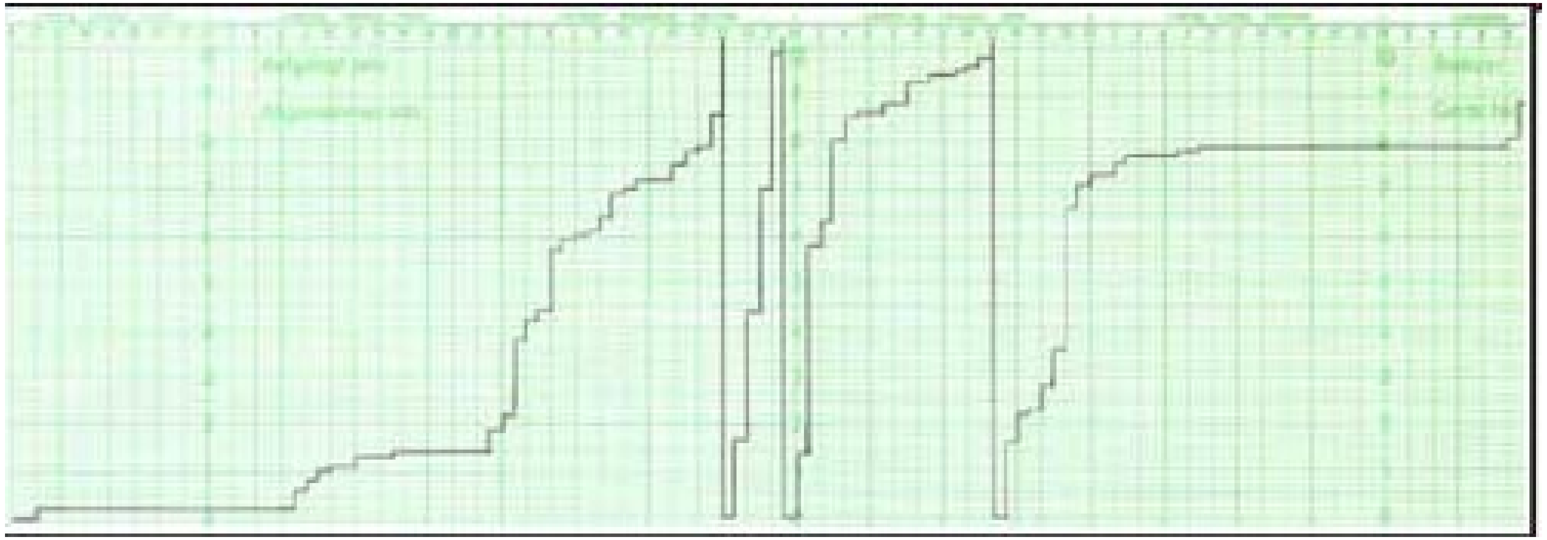
Standard brass rain gauge (Casella)



Components of a pole mounted standard rain gauge

Penakar Hujan Jenis Timbangan

Tipe timbangan (*weighing bucket*) dapat merekam jumlah kumulatif hujan secara kontinyu. Alat ini tidak dilengkapi dengan sistem pengurasan otomatik.



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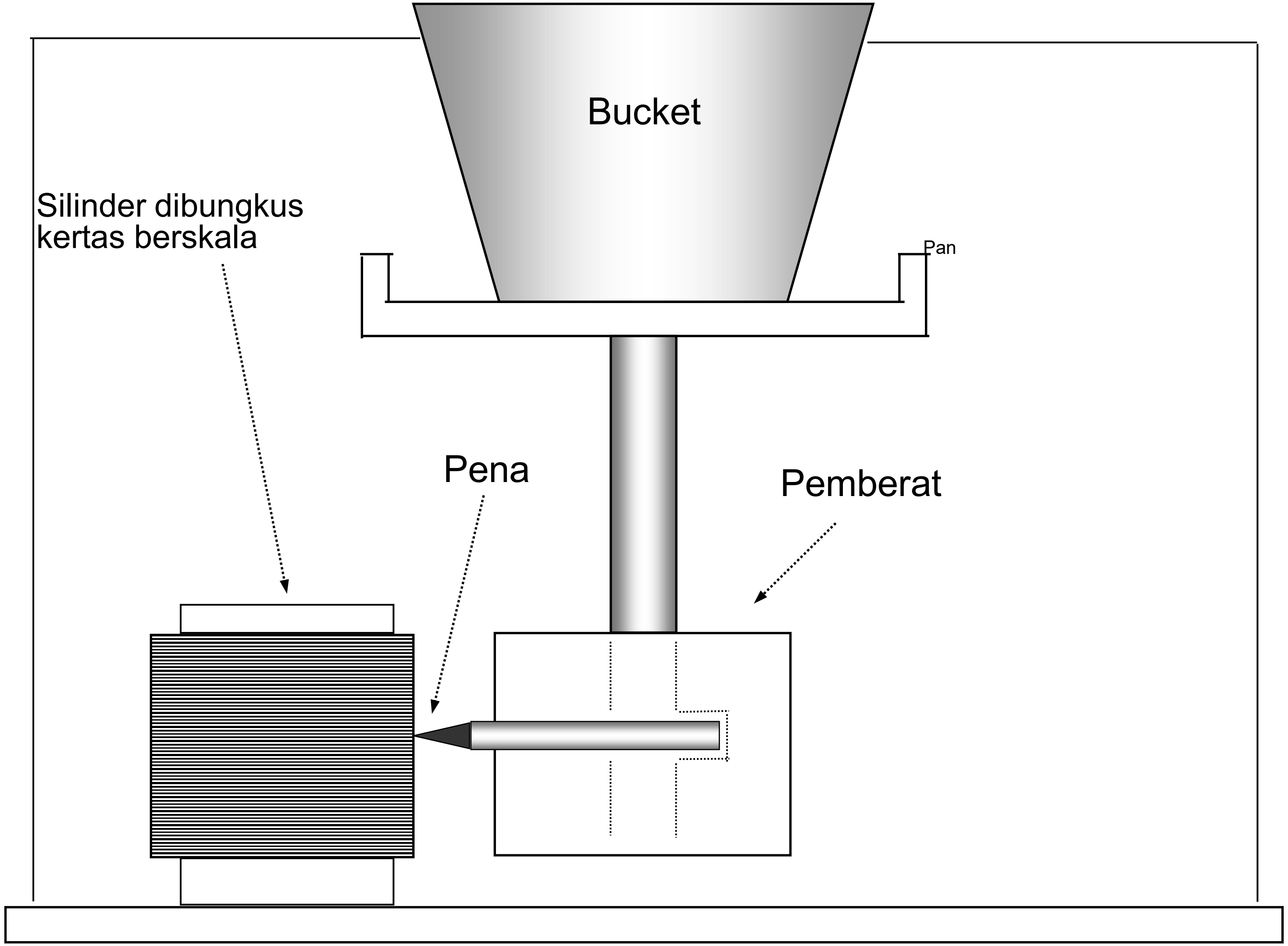
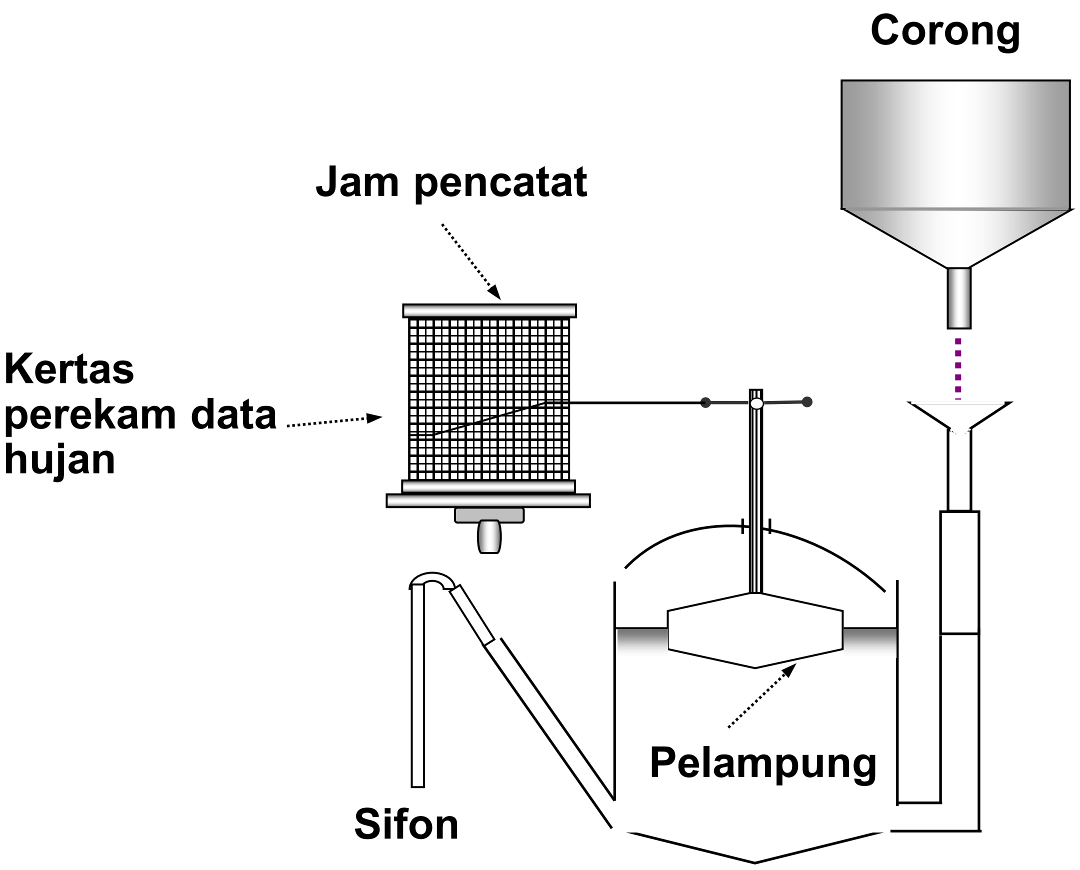


Alat Penakar Hujan Jenis Pelampung

Prinsip mekanisme kerja alat penakar hujan otomatis tipe ketiga yaitu *float* adalah dengan memanfaatkan gerakan naik pelampung dalam bejana akibat tertampungnya curah hujan. Pelampung ini berhubungan dengan sistem pena perekam di atas kertas berskala yang menghasilkan grafik rekaman data hujan. Alat ini dilengkapi dengan sistem pengurasan otomatis, yaitu pada saat air hujan yang tertampung telah mencapai kapasitas *receiver*nya akan dikeluarkan dari bejana dan pena akan kembali pada posisi dasar kertas rekaman data hujan.



Syarat teknis Penempatan dan pemasangan alat pada stasiun hidrologi

- Penakar hujan ditempatkan pada lokasi sedemikian sehingga kecepatan angin di tempat tersebut sekecil mungkin dan terhindar dari pengaruh penangkapan air hujan oleh benda lain di sekitar alat penakar hujan.
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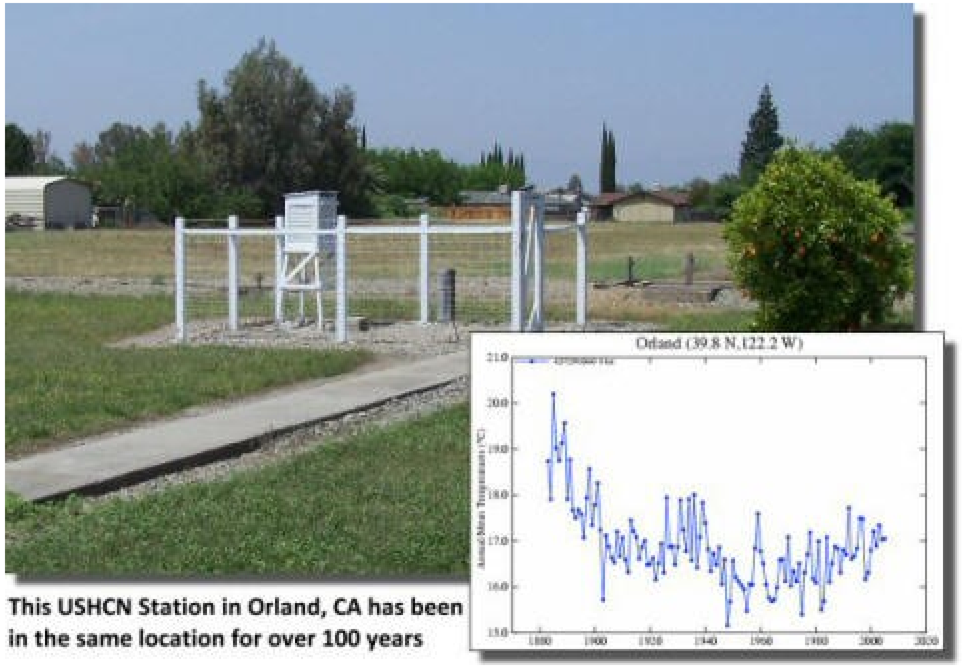
Orland (39.8 N,122.2 W)
This USHCN Station in Orland, CA has been in the same location for over 100 years


BMKG
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Automatic Weather Station

Metode Perhitungan Hujan Rerata

- Dalam analisis hidrologi sering diperlukan untuk menentukan hujan rerata pada daerah tersebut.
- Terdapat 3 metode :
  - Aritmatik
  - Poligon Thiessen
  - Isohiet

1. Metode rerata aritmatik (aljabar)

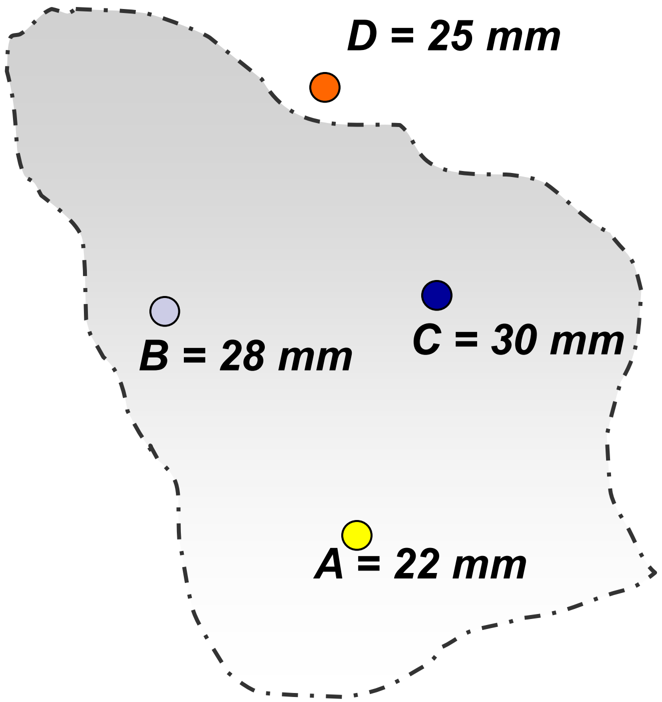
- Metode ini adalah metode yang paling sederhana. Pengukuran dengan metode ini dilakukan dengan merata-ratakan hujan di seluruh DAS. Stasiun hujan yang digunakan untuk menghitung dengan metode ini adalah yang berada di dalam DAS, akan tetapi stasiun yang berada di luar DAS dan jaraknya cukup berdekatan masih bisa diperhitungkan. Metode aljabar ini memberikan hasil yang tidak teliti, metode ini memberikan hasil yang cukup baik jika penyebaran hujan merata, serta hujan tidak terlalu bervariasi.
- Hujan DAS dengan cara ini dapat diperoleh dengan persamaan:

$$\bar{p} = \frac{\sum_{i=1}^{n} p_i}{n} \qquad \bar{p} = \frac{p_1 + p_2 + p_3 + \ldots + p_n}{n}$$

- dengan:

  p = hujan rerata di suatu DAS
  p_i = hujan di tiap-tiap stasiun
  n = jumlah stasiun

Contoh Ilustrasi



Hitung hujan rerata dengan metode aljabar!

$$\bar{p} = \frac{p_1 + p_2 + p_3 + \ldots\ldots + p_n}{n}$$

$$\bar{p} = \frac{p_A + p_B + p_C}{3}$$

$$\bar{p} = \frac{22 + 28 + 30}{3}$$

$$\bar{p} = 26{,}67\,mm$$

Jika stasiun D di luar DAS ikut diperhitungkan maka:

$$\bar{p} = \frac{22 + 28 + 30 + 25}{4} = 26{,}25\,mm$$

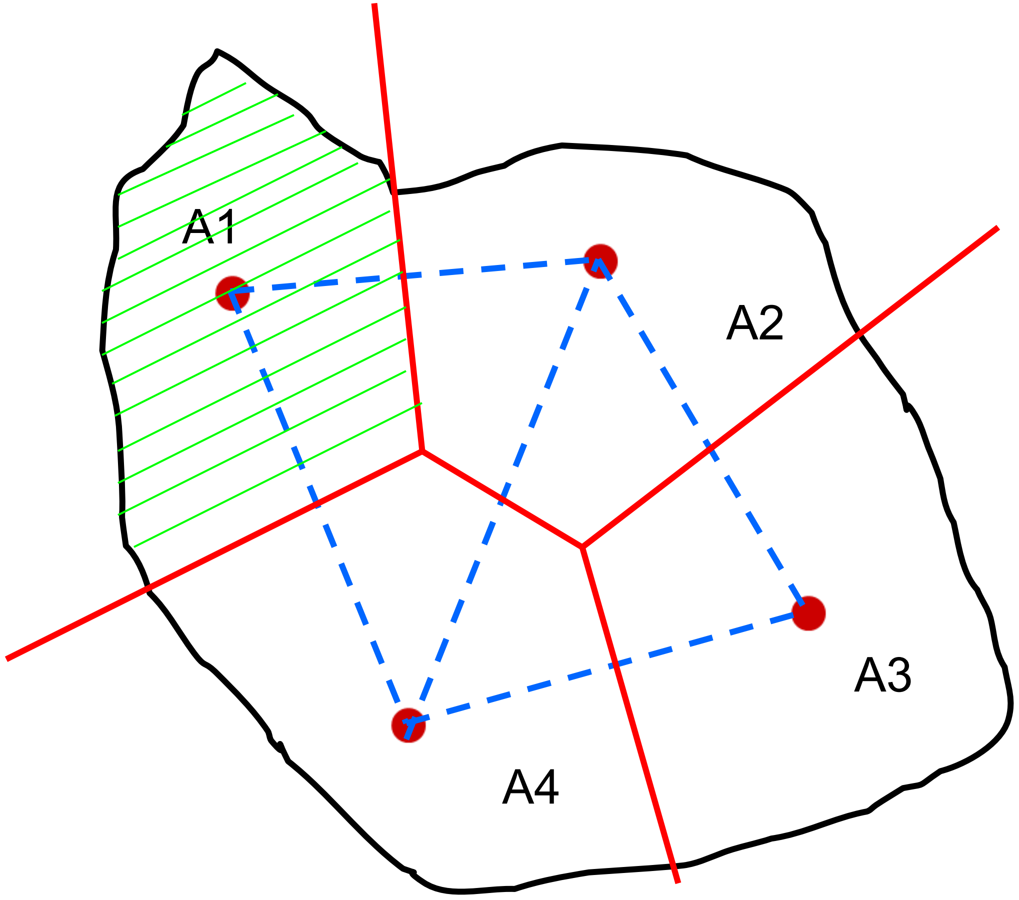
2. Metode Thiessen

- Metode ini digunakan untuk menghitung bobot masing-masing stasiun yang mewakili luasan di sekitarnya. Metode ini digunakan bila penyebaran hujan di daerah yang ditinjau tidak merata.

PROSEDUR HITUNGAN METODE POLIGON THIESSEN

Hitungan poligon Thiessen dilakukan dengan cara:

a. Stasiun hujan digambar pada peta daerah yang ditinjau.
b. Stasiun-stasiun tersebut dihubungkan dengan garis lurus, sehingga akan didapatkan bentuk segitiga.
c. Tiap-tiap sisi segitiga dibuat garis berat sehingga saling bertemu dan membentuk suatu poligon yang mengelilingi tiap stasiun. Tiap stasiun mewakili luasan yang dibentuk oleh poligon, sedangkan untuk stasiun yang berada di dekat batas daerah, garis batas daerah membentuk batas tertutup dari poligon.
d. Luas tiap poligon diukur, kemudian dikalikan dengan kedalaman hujan di tiap poligon. Hasil jumlah hitungan tersebut dibagi dengan total luas daerah yang ditinjau.


A1
A2
A3
A4

Prosedur hitungan ini dijelaskan pada persamaan dan gambar berikut ini.

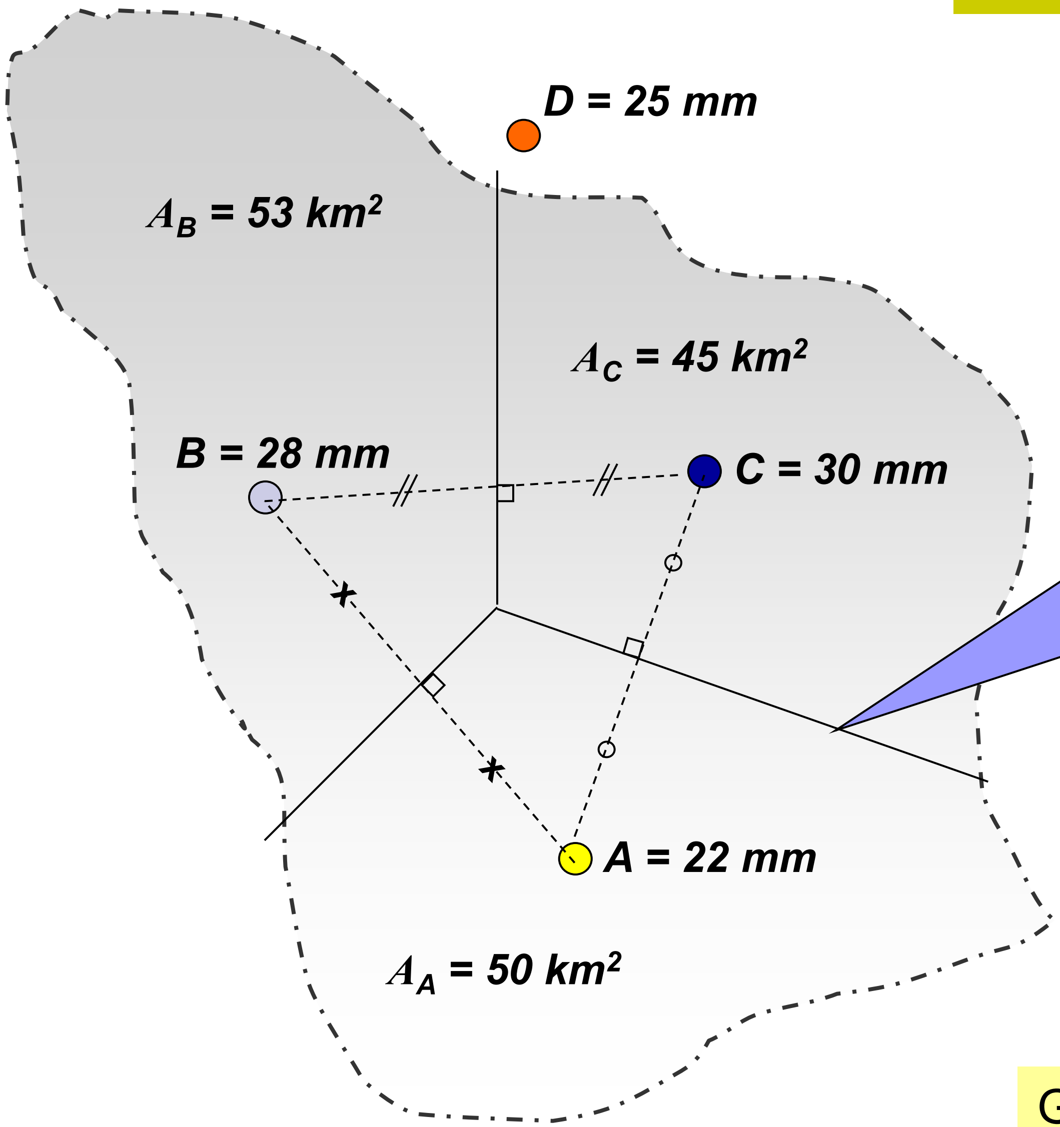
$$\overline{P} = \frac{A_1.P_1 + A_2.P_2 + ...+ A_n.P_n}{A_{total}}$$

$$\overline{P} = \frac{A_1.P_1 + A_2.P_2 + A_3.P_3 + ...+ A_n.P_n}{A_1 + A_2 + A_3 + ...+ A_n}$$

Dimana:

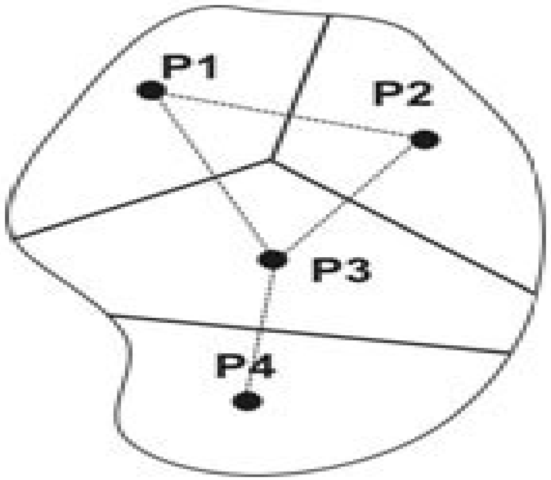
- P = curah hujan rata-rata,
- $P_1,..., P_n$ = curah hujan pada setiap setasiun,
- $A_1,..., A_n$ = luas yang dibatasi tiap poligon.

Contoh Ilustrasi



Garis ini membagi sisi segitiga menjadi 2 bagian sama panjang (di tengah-tengah) dan tegak lurus terhadapnya.

Gambar tidak berskala, luas bagian dan tinggi hujan hanya merupakan perumpamaan



$$P = \frac{1A1 + P2A2 + P3A3 + P4A}{A1 + A2 + A3 + A4} \quad \ldots (2)$$

Keterangan :

$\bar{P}$ = hujan rata-rata

P1, P2, P3, P4 = tebal hujan pada stasiun 1,2,3,4

A1, A2, A3, A4 = luas wilayah yang diwakili oleh stasiun 1,2,3,4.

3. Metode Isohiet

- Pada prinsipnya isohiet adalah garis yang menghubungkan titik-titik dengan tinggi/kedalaman hujan yang sama, Kesulitan dari penggunaan metode ini adalah jika jumlah stasiun di dalam dan sekitar DAS terlalu sedikit. Hal tersebut akan mengakibatkan kesulitan dalam menginterpolasi.

Metode pembuatan garis Isohiet sebagai berikut:

- Pada peta yang ditinjau, digambarkan lokasi daerah hujan dan kedalaman hujan.
- Di stasiun hujan yang saling berdampingan dinilai kedalaman hujannya dan dibuat interpolasinya. Kemudian hasil interpolasi yang mewakili kedalaman hujan yang sama dihubungkan satu sama lain.
- Luas daerah diantara 2 garis isohiet diukur luasnya, dan dikalikan dengan nilai rerata di kedua garis isohiet. Kemudian jumlah dari hasil hitungan tersebut dibagi dengan total luasan daerah yang ditinjau.

Hujan DAS menggunakan Isohiet dapat dihitung dengan persamaan:

$$\bar{p} = \frac{\sum_{i=1}^{n} A_i \frac{I_i + I_{i+1}}{2}}{\sum_{i}^{n} A_i}$$

$$\bar{p} = \frac{A_1 \frac{I_1 + I_2}{2} + A_2 \frac{I_2 + I_3}{2} + \ldots.. + A_n \frac{I_n + I_{n+1}}{2}}{A_1 + A_2 + \ldots.. + A_n}$$

Dengan:

p = hujan rerata kawasan
A_i = luasan dari titik i
I_i = garis isohiet ke i

KONDISI DAN SIFAT DATA

- Data hujan yang baik diperlukan dalam melakukan analisis hidrologi, namun untuk mendapatkan data yang berkualitas biasanya tidak mudah. Data hujan hasil pencatatan yang tersedia biasanya dalam kondisi tidak menerus. Apabila terputusnya rangkaian data hanya beberapa saat kemungkinan tidak menimbulkan masalah tetapi untuk kurun waktu yang lama tentu akan menimbulkan masalah di dalam melakukan analisis.

- Dalam hal ini perlu dilihat kepentingan atau sasaran dari perencanaan drainase yang bersangkutan.

Melengkapi Data

- Jika ada data hilang atau tidak lengkap

$$r = \tfrac{1}{3}\left(\frac{R}{R_A} r_A + \frac{R}{R_B} r_B + \frac{R}{R_C} r_C\right)$$

dengan:
R = curah hujan rata-rata setahun di tempat pengamatan R datanya harus lengkap
r_A = curah hujan ditempat pengamatan RA
RA = curah hujan rata-rata setahun di A

"Dan Dialah yang meniupkan angin sebagai pembawa berita gembira di muka kedatangan rahmatNya (hujan), hingga apabila angin itu telah membawa awan mendung, Kami halau ke suatu daerah yang tandus, lalu Kami turunkan hujan di daerah itu. Maka Kami keluarkan dengan sebab hujan ini pelbagai macam buah-buahan. Seperti itulah Kami membangkitkan orang-orang yang telah mati, supaya kamu mengambil pelajaran."

Surat 7 (Al A'Raaf),ayat 57

TERIMAKASIH

MATERI – 2

HUJAN (PRESIPITASI)

Disampaikan Oleh :
Ir. Rahardjo Samiono M.T
Muhamad Komarudin S.Si., M.Si

Pengertian Hujan

Hujan adalah peristiwa turunnya butir-butir air dari langit ke permukaan bumi. Hujan juga merupakan siklus air di bumi.

Curah Hujan

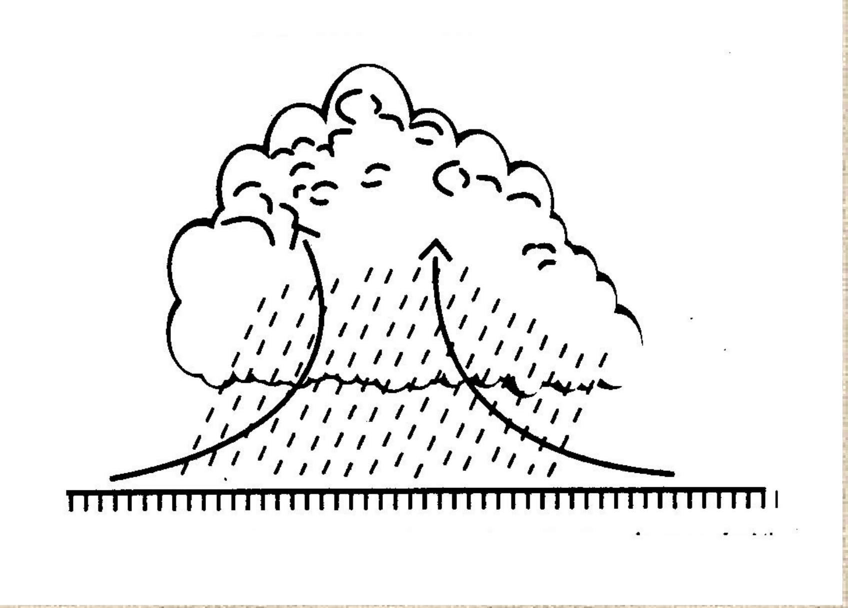
Curah hujan adalah jumlah air yang jatuh di permukaan tanah datar selama periode tertentu yang diukur dengan satuan tinggi (mm).

Tipe Hujan

- Hujan terjadi karena udara basah yang naik ke atmosfer mengalami pendinginan sehingga terjadi proses kondensasi.
- Naiknya udara ke atas dapat terjadi secara siklonik, orografik dan konvektif.

Hujan Konvektif

- Hujan jenis ini biasanya terjadi sebagai hujan dengan intensitas yang tinggi, akibat massa udara yang terangkat ke atas oleh pemanasan lahan. Hujan jenis ini biasanya terjadi di daerah yang relatif luas dan bergerak sesuai dengan pergerakan angin.



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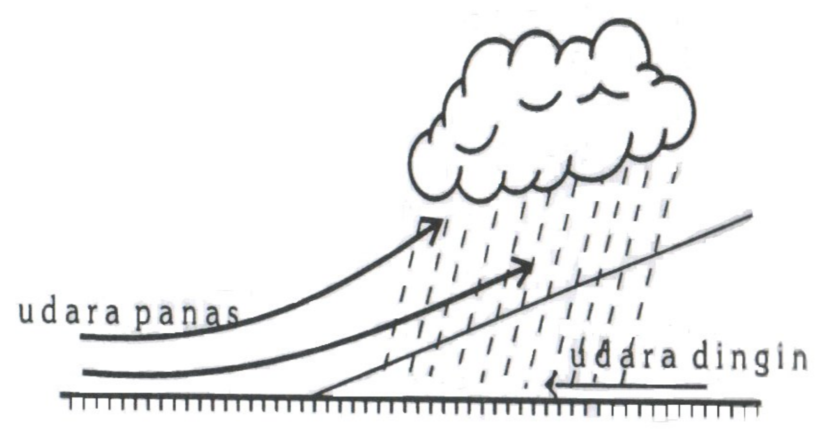

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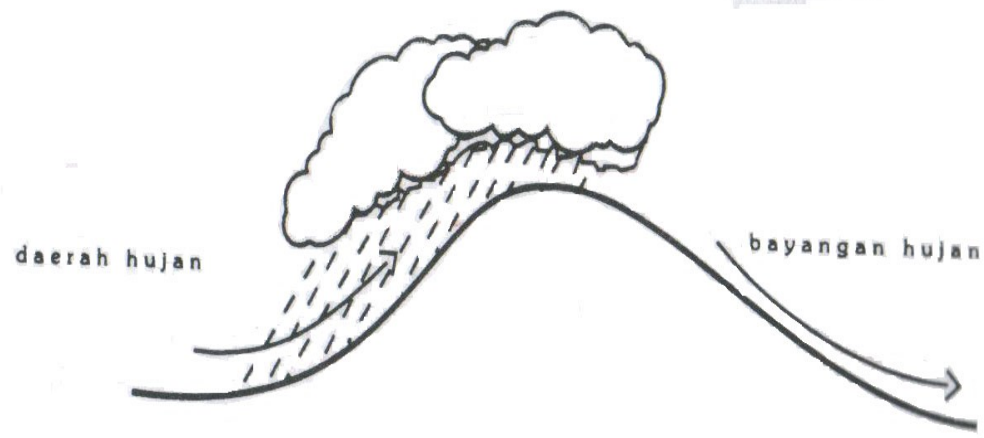

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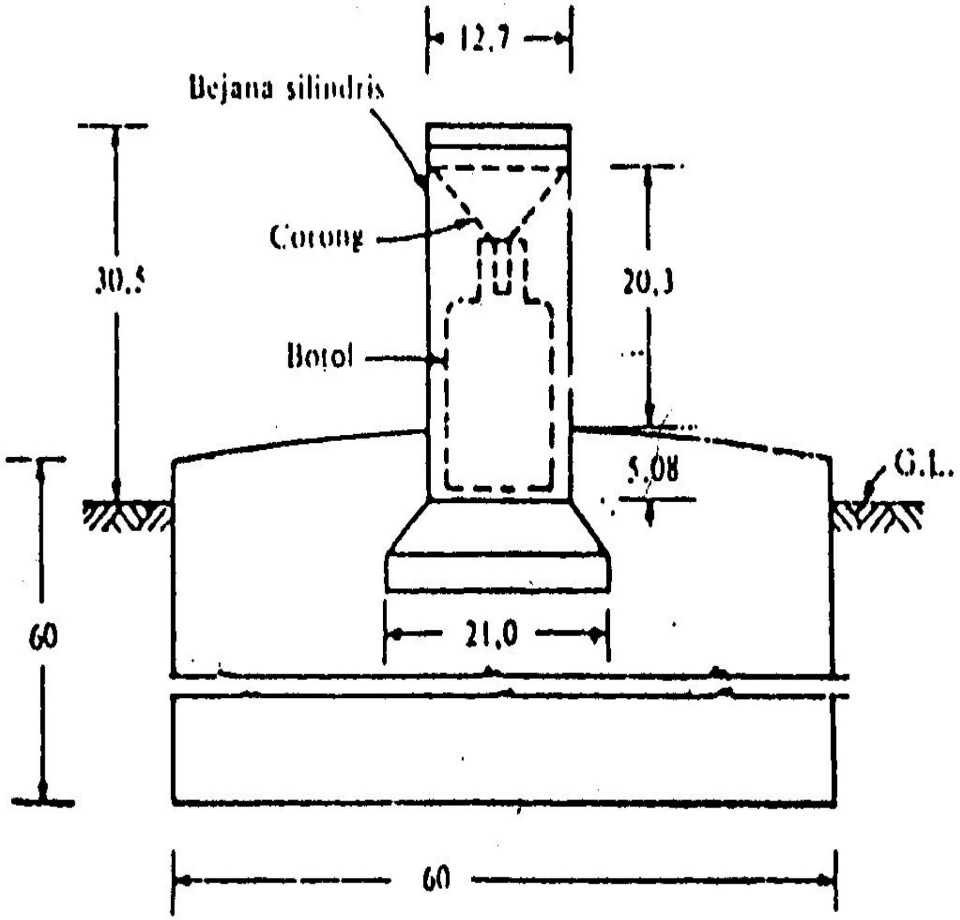

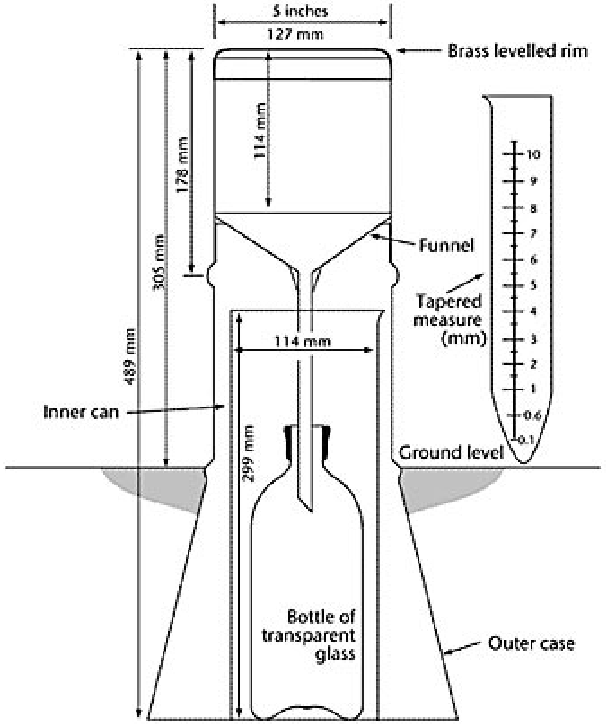

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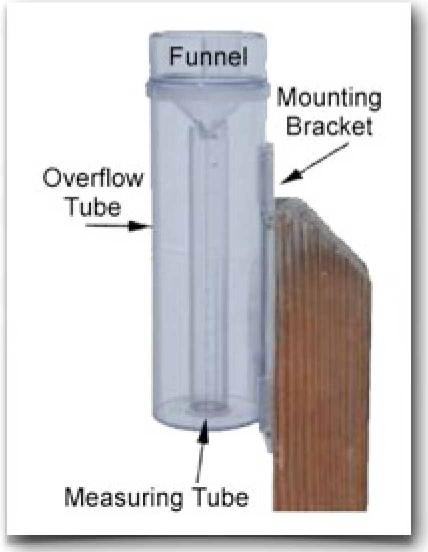

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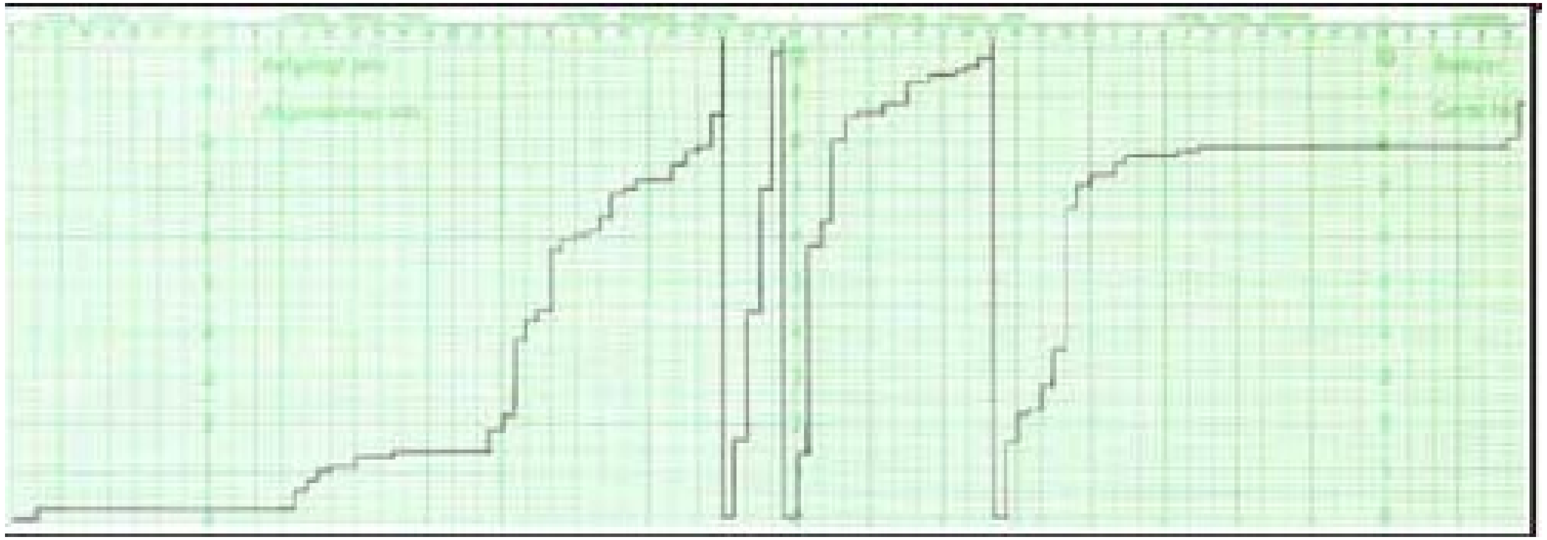

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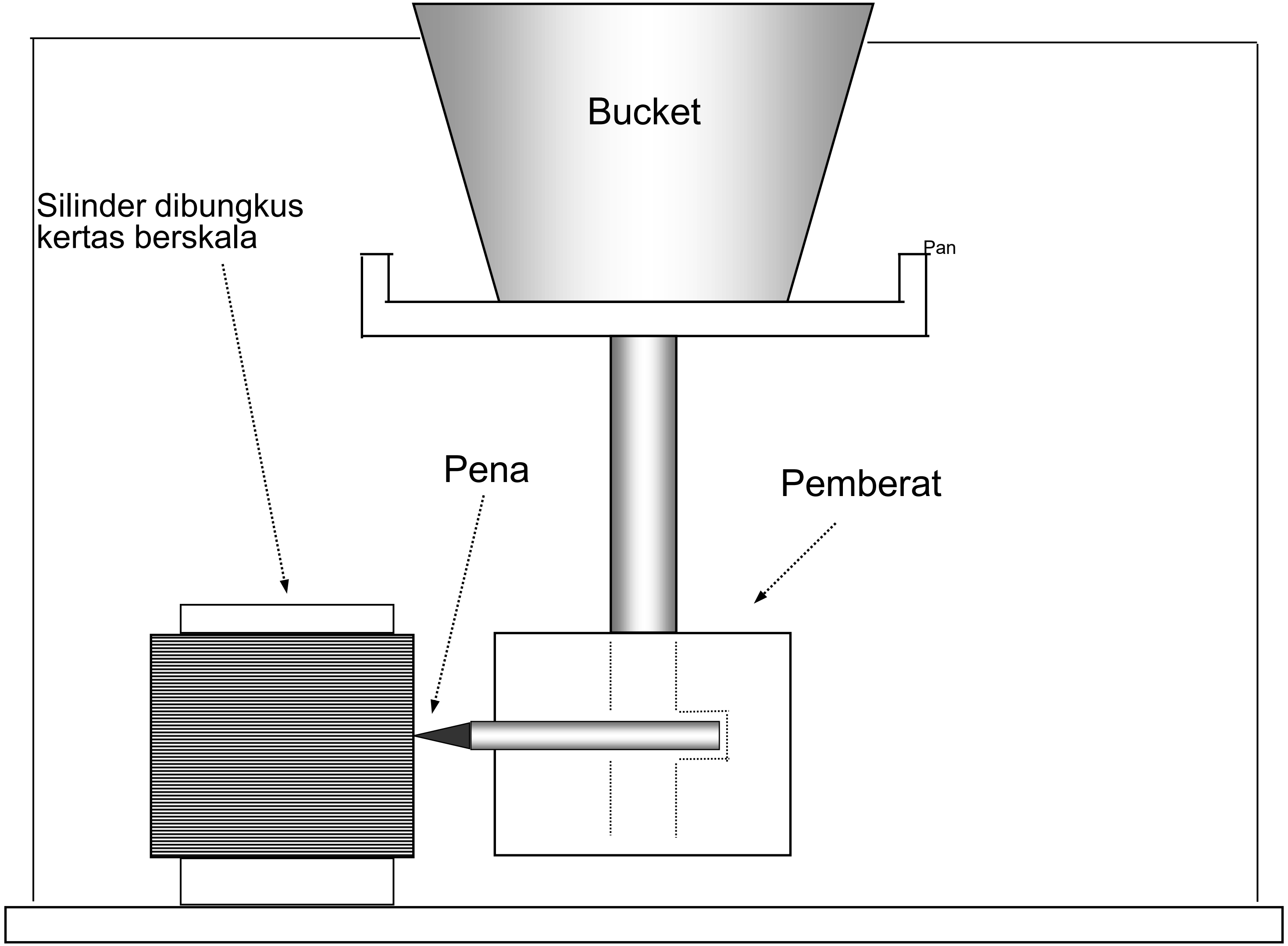

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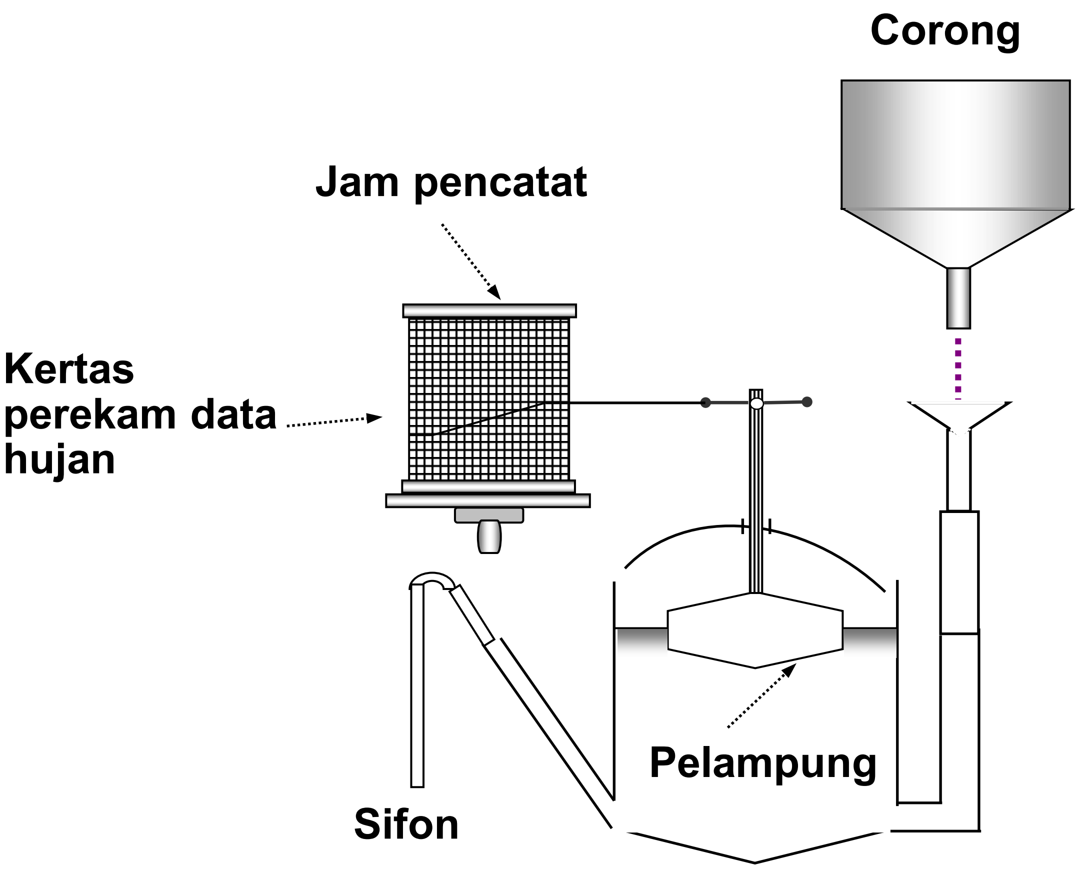

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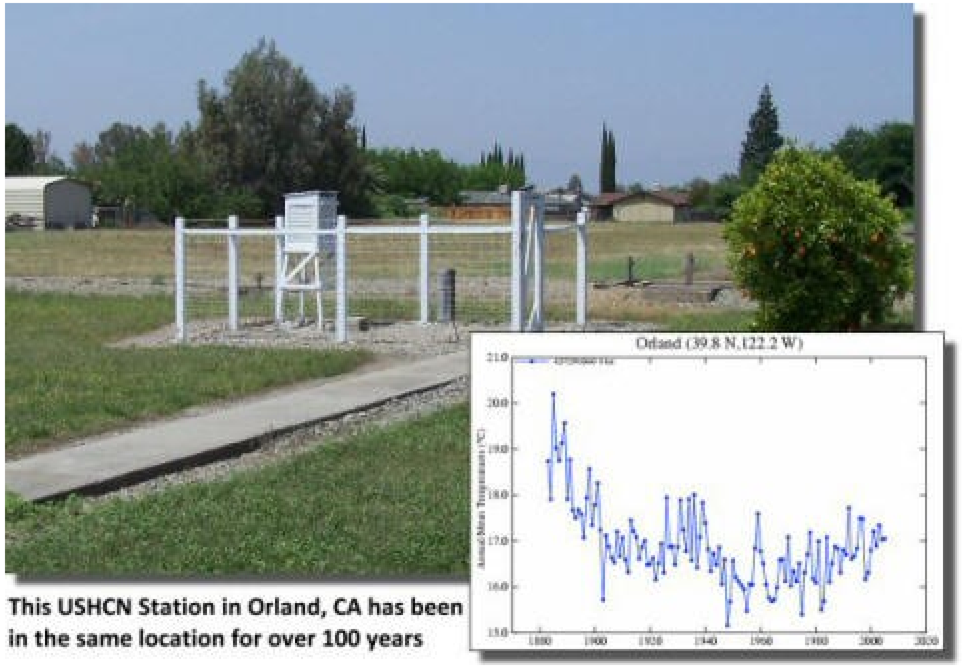
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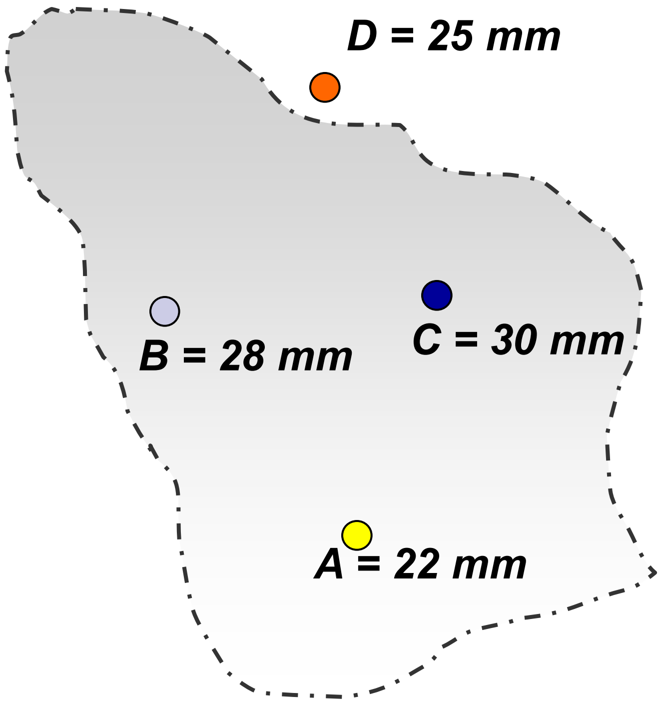
$$\bar{p} = \frac{\sum_{i=1}^{n} p_i}{n} \qquad \bar{p} = \frac{p_1 + p_2 + p_3 + \ldots\ldots + p_n}{n}$$

- dengan:

p	= hujan rerata di suatu DAS
p_i	= hujan di tiap-tiap stasiun
n	= jumlah stasiun

Contoh Ilustrasi



Hitung hujan rerata dengan metode aljabar!

$$\overline{p} = \frac{p_1 + p_2 + p_3 + \ldots\ldots + p_n}{n}$$

$$\overline{p} = \frac{p_A + p_B + p_C}{3}$$

$$\overline{p} = \frac{22 + 28 + 30}{3}$$

$$\overline{p} = 26{,}67 mm$$

Jika stasiun D di luar DAS ikut diperhitungkan maka:

$$\overline{p} = \frac{22 + 28 + 30 + 25}{4} = 26{,}25 mm$$

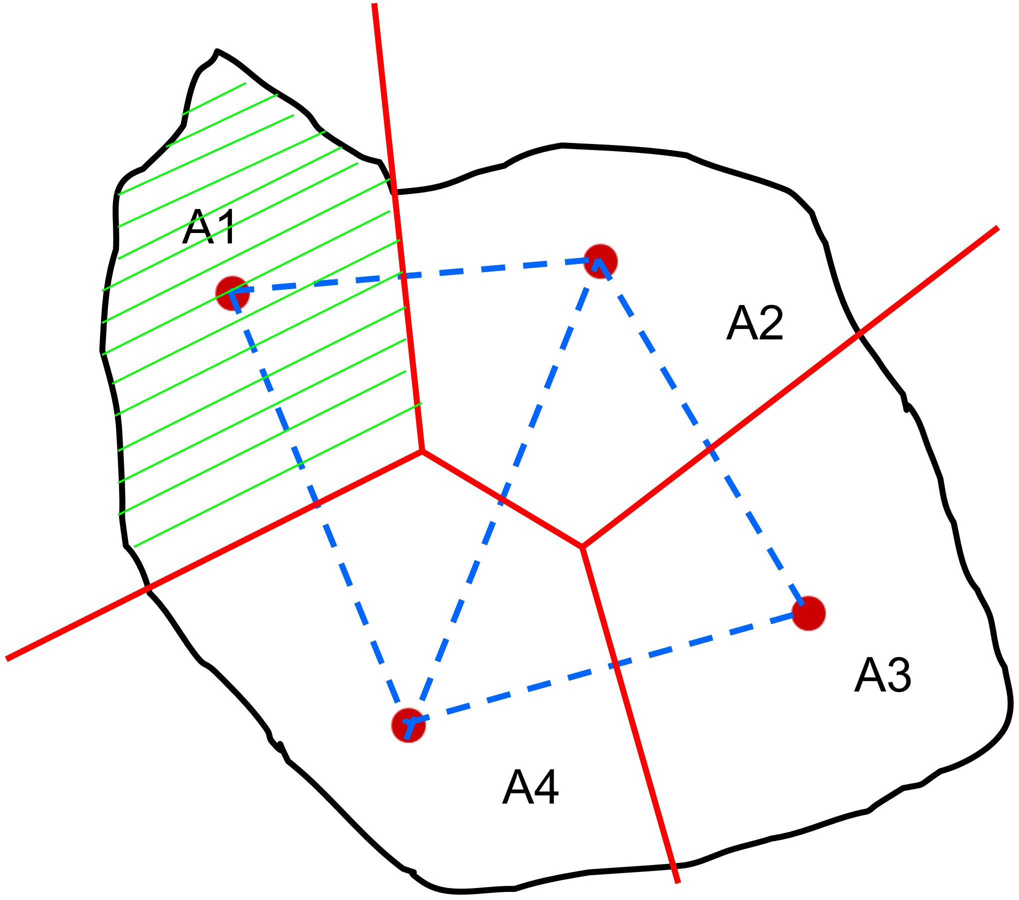
2. Metode Thiessen

- Metode ini digunakan untuk menghitung bobot masing-masing stasiun yang mewakili luasan di sekitarnya. Metode ini digunakan bila penyebaran hujan di daerah yang ditinjau tidak merata.

PROSEDUR HITUNGAN METODE POLIGON THIESSEN

Hitungan poligon Thiessen dilakukan dengan cara:

a. Stasiun hujan digambar pada peta daerah yang ditinjau.
b. Stasiun-stasiun tersebut dihubungkan dengan garis lurus, sehingga akan didapatkan bentuk segitiga.
c. Tiap-tiap sisi segitiga dibuat garis berat sehingga saling bertemu dan membentuk suatu poligon yang mengelilingi tiap stasiun. Tiap stasiun mewakili luasan yang dibentuk oleh poligon, sedangkan untuk stasiun yang berada di dekat batas daerah, garis batas daerah membentuk batas tertutup dari poligon.
d. Luas tiap poligon diukur, kemudian dikalikan dengan kedalaman hujan di tiap poligon. Hasil jumlah hitungan tersebut dibagi dengan total luas daerah yang ditinjau.


A1
A2
A3
A4

Prosedur hitungan ini dijelaskan pada persamaan dan gambar berikut ini.

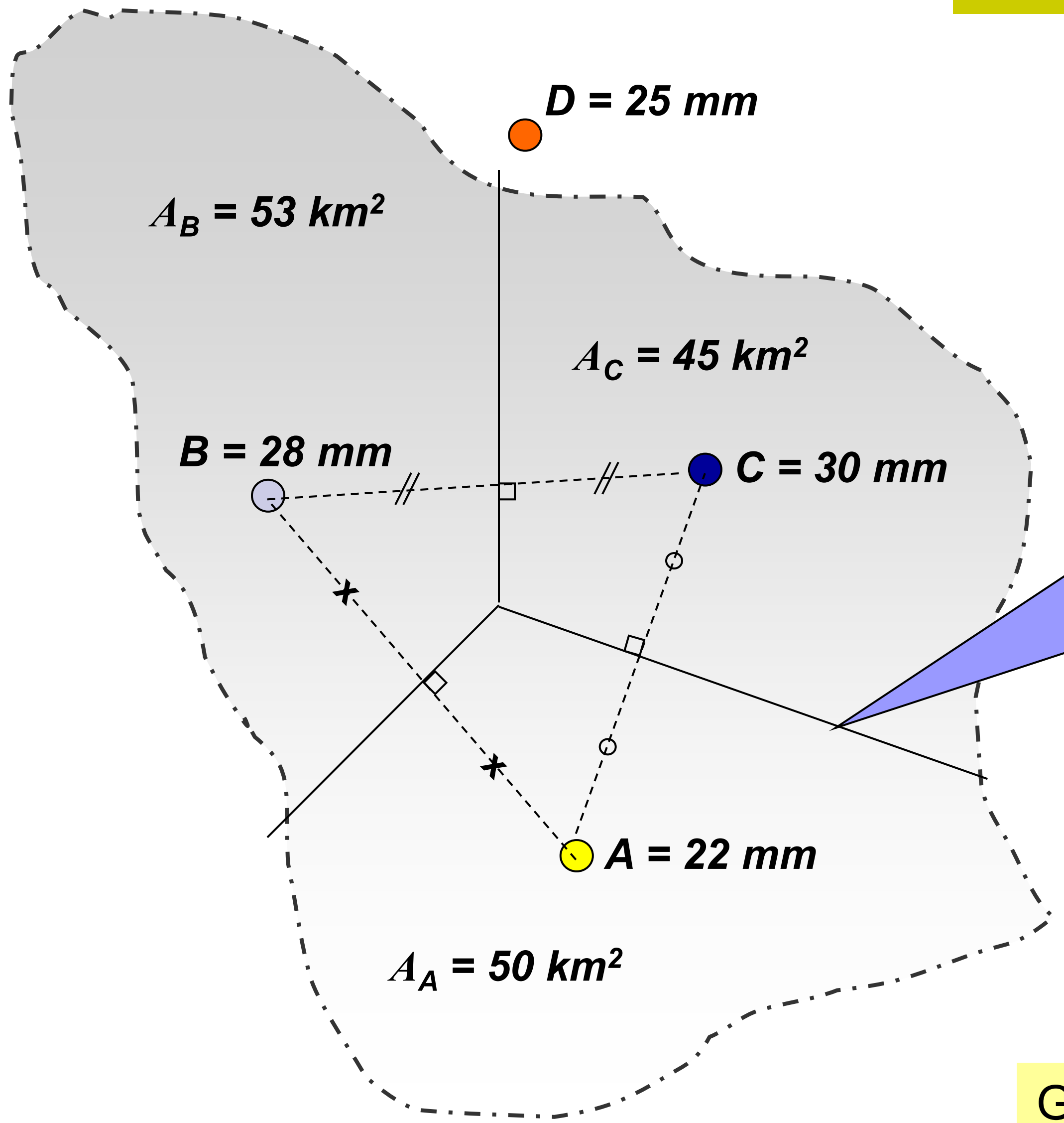
$$\overline{P} = \frac{A_1.P_1 + A_2.P_2 + ...+ A_n.P_n}{A_{total}}$$

$$\overline{P} = \frac{A_1.P_1 + A_2.P_2 + A_3.P_3 + ...+ A_n.P_n}{A_1 + A_2 + A_3 + ...+ A_n}$$

Dimana:

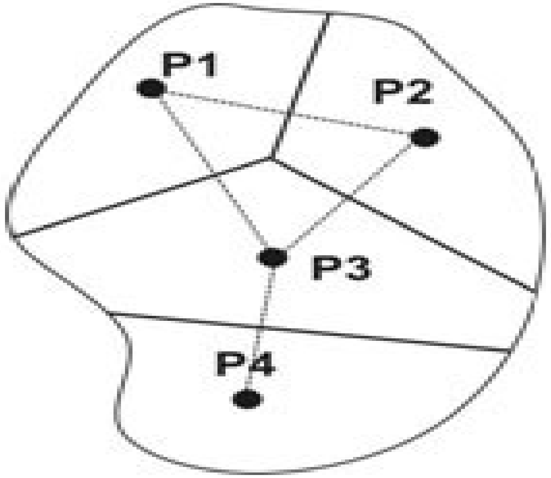
- P = curah hujan rata-rata,
- $P_1,..., P_n$ = curah hujan pada setiap setasiun,
- $A_1,..., A_n$ = luas yang dibatasi tiap poligon.

Contoh Ilustrasi



Garis ini membagi sisi segitiga menjadi 2 bagian sama panjang (di tengah-tengah) dan tegak lurus terhadapnya.

Gambar tidak berskala, luas bagian dan tinggi hujan hanya merupakan perumpamaan



$$P = \frac{1A1 + P2A2 + P3A3 + P4A}{A1 + A2 + A3 + A4} \quad \ldots (2)$$

Keterangan :

$\overline{P}$ = hujan rata-rata

P1, P2, P3, P4 = tebal hujan pada stasiun 1,2,3,4

A1, A2, A3, A4 = luas wilayah yang diwakili oleh stasiun 1,2,3,4.

3. Metode Isohiet

- Pada prinsipnya isohiet adalah garis yang menghubungkan titik-titik dengan tinggi/kedalaman hujan yang sama, Kesulitan dari penggunaan metode ini adalah jika jumlah stasiun di dalam dan sekitar DAS terlalu sedikit. Hal tersebut akan mengakibatkan kesulitan dalam menginterpolasi.

Metode pembuatan garis Isohiet sebagai berikut:

- Pada peta yang ditinjau, digambarkan lokasi daerah hujan dan kedalaman hujan.
- Di stasiun hujan yang saling berdampingan dinilai kedalaman hujannya dan dibuat interpolasinya. Kemudian hasil interpolasi yang mewakili kedalaman hujan yang sama dihubungkan satu sama lain.
- Luas daerah diantara 2 garis isohiet diukur luasnya, dan dikalikan dengan nilai rerata di kedua garis isohiet. Kemudian jumlah dari hasil hitungan tersebut dibagi dengan total luasan daerah yang ditinjau.

Hujan DAS menggunakan Isohiet dapat dihitung dengan persamaan:

$$\bar{p} = \frac{\sum_{i=1}^{n} A_i \frac{I_i + I_{i+1}}{2}}{\sum_{i}^{n} A_i}$$

$$\bar{p} = \frac{A_1 \frac{I_1 + I_2}{2} + A_2 \frac{I_2 + I_3}{2} + \ldots\ldots + A_n \frac{I_n + I_{n+1}}{2}}{A_1 + A_2 + \ldots\ldots + A_n}$$

Dengan:

p = hujan rerata kawasan

A_i = luasan dari titik i

I_i = garis isohiet ke i

KONDISI DAN SIFAT DATA

- Data hujan yang baik diperlukan dalam melakukan analisis hidrologi, namun untuk mendapatkan data yang berkualitas biasanya tidak mudah. Data hujan hasil pencatatan yang tersedia biasanya dalam kondisi tidak menerus. Apabila terputusnya rangkaian data hanya beberapa saat kemungkinan tidak menimbulkan masalah tetapi untuk kurun waktu yang lama tentu akan menimbulkan masalah di dalam melakukan analisis.

- Dalam hal ini perlu dilihat kepentingan atau sasaran dari perencanaan drainase yang bersangkutan.

Melengkapi Data

- Jika ada data hilang atau tidak lengkap

$$r = \frac{1}{3}\left(\frac{R}{R_A} r_A + \frac{R}{R_B} r_B + \frac{R}{R_C} r_C \right)$$

dengan:
R = curah hujan rata-rata setahun di tempat pengamatan R datanya harus lengkap
r_A = curah hujan ditempat pengamatan RA
RA = curah hujan rata-rata setahun di A

"Dan Dialah yang meniupkan angin sebagai pembawa berita gembira di muka kedatangan rahmatNya (hujan), hingga apabila angin itu telah membawa awan mendung, Kami halau ke suatu daerah yang tandus, lalu Kami turunkan hujan di daerah itu. Maka Kami keluarkan dengan sebab hujan ini pelbagai macam buah-buahan. Seperti itulah Kami membangkitkan orang-orang yang telah mati, supaya kamu mengambil pelajaran."

Surat 7 (Al A'Raaf),ayat 57

TERIMAKASIH

MATERI – 3

EVAPORASI DAN INFILTRASI

Disampaikan Oleh :
Ir. Rahardjo Samiono M.T
Muhamad Komarudin S.Si., M.Si

EVAPORASI

Definisi

Penguapan adalah proses berubahnya bentuk zat cair (air) menjadi gas (uap air) dan masuk ke atmosfer.

Di dalam hidrologi, penguapan dibagi menjadi dua:

1. **Evaporasi (Ep)**

   adalah penguapan yang terjadi dari permukaan air (seperti laut, danau, sungai), permukaan tanah (genangan di atas tanah dan penguapan dari permukaan air tanah yang dekat dengan permukaan tanah), dan permukaan tanaman (intersepsi). Intersepsi adalah penguapan yang berasal dari air hujan yang berada pada permukaan daun, ranting dan badan tanaman.

2. **Transpirasi (Et)**

   adalah penguapan melalui tanaman, dimana air tanah diserap oleh akar tanaman yang kemudian dialirkan melalui batang sampai ke permukaan daun dan menguap menuju atmosfer.

Oleh karena sulitnya membedakan antara penguapan dari lahan air, tanah dan tanaman, maka biasanya evaporasi dan transpirasi dicakup menjadi satu yaitu **evapotranspirasi**.

Evapotranspirasi dapat didefinisikan sebagai penguapan yang terjadi di permukaan lahan, yang meliputi permukaan tanah dan tanaman yang tumbuh di permukaan lahan tersebut. Apabila ketersediaan air (lengas tanah) tak terbatas, maka evapotranspirasi yang terjadi disebut evapotranspirasi potensial (ETP). Akan tetapi pada umumnya ketersediaan air di permukaan tidak tak terbatas, sehingga evapotranspirasi terjadi dengan laju lebih kecil dari evapotranspirasi potensial. Evapotanspirasi yang terjadi sebenarnya terjadi di suatu daerah disebut evapotranspirasi nyata.

Faktor-faktor yang Mempengaruhi Evapotranspirasi

1. Radiasi matahari

   Radiasi matahari merupakan sumber utama panas. Hal tersebut mempengaruhi jumlah evapotranspirasi di atas permukaan bumi yang tergantung pada garis lintang dan musim.

2. Temperatur

   Semakin tinggi temperatur, semakin besar kemampuan udara untuk menyerap uap air. Selain itu, semakin tinggi temperatur, energi kinetik molekul air meningkat, sehingga molekul air semakin banyak yang berpindah ke lapis udara di atasnya dalam bentuk uap air.

3. Kelembaban

Perbedaan tekanan uap menyebabkan terjadinya penguapan. Apabila jumlah uap air yang masuk ke udara semakin banyak, tekanan uap airnya juga semakin tinggi. Akibatnya perbedaan tekanan uap semakin kecil, sehingga menyebabkan berkurangnya laju penguapan. Apabila udara di atas permukaan air sudah jenuh uap air, tekanan udara telah mencapai tekanan uap jenuh, di mana pada saat itu penguapan terhenti.

4. Angin

Apabila proses evaporasi terus berlangsung, udara akan menjadi jenuh terhadap uap air dan evaporasi akan terhenti. Agar proses penguapan dapat berjalan terus, lapisan udara yang telah jenuh harus diganti dengan udara kering. Penggantian tersebut dapat terjadi apabila ada angin. Di daerah terbuka dan banyak angin, penguapan akan lebih besar daripada di daerah yang terlindung dan udara diam.

Pengukuran Evaporasi

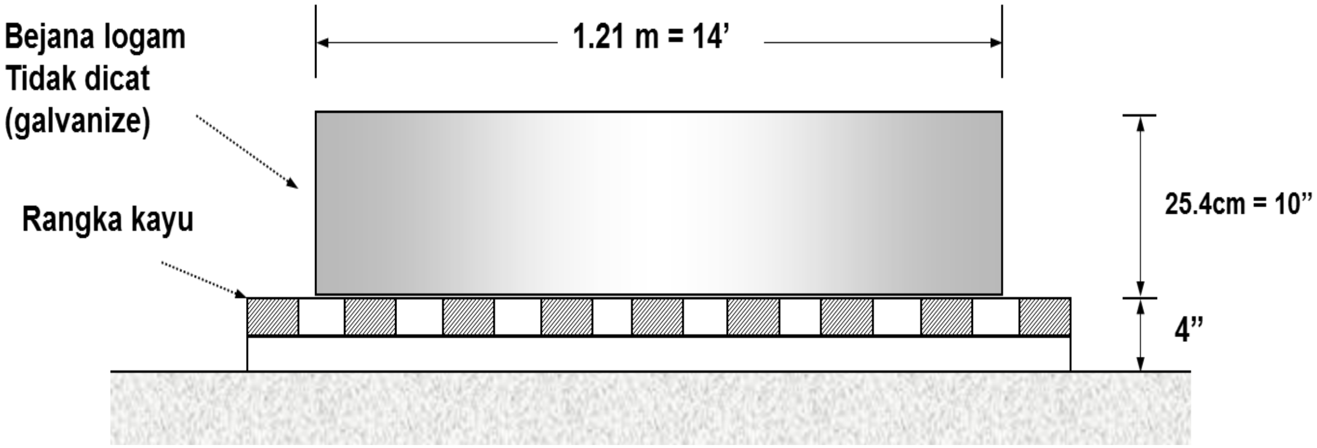
Besarnya evaporasi dapat diperkirakan dengan pendekatan teoritis maupun dengan pengukuran langsung. Cara pertama memerlukan banyak data meteorologi dan data penunjang lain yang tidak selalu mudah didapatkan. Oleh karena itu pengukuran langsung di lapangan sering dilakukan untuk keperluan analisis secara lebih praktis.

1. Atmometer

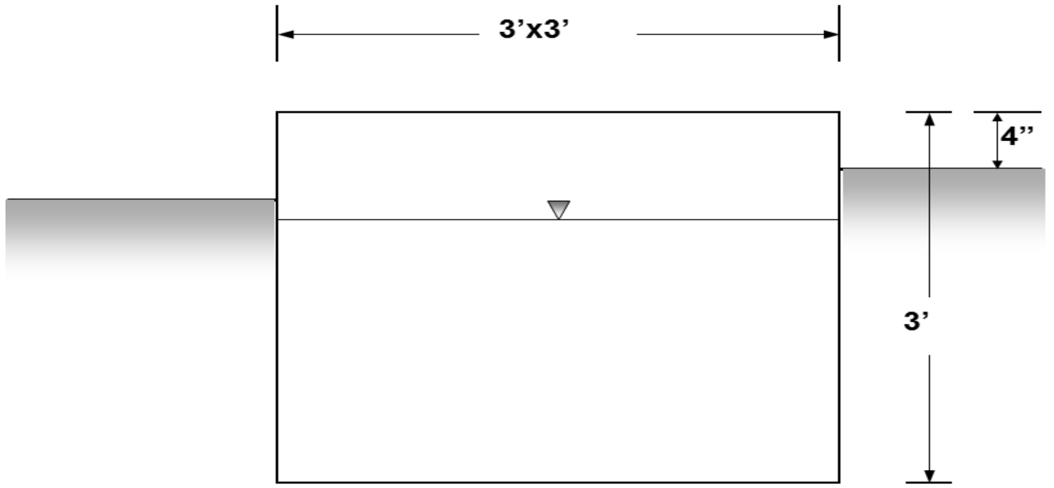
Alat pengukur evaporasi ini cukup sederhana, berupa bejana berpori yang diisi air. Besarnya penguapan dalam jangka waktu tertentu, misalnya harian didapatkan dari nilai selisih pembacaan sebelum dan sesudah percobaan. Beberapa jenis atmometer antara lain Piche, Livingstone dan Black Bellani.

2. Evaporation Pan

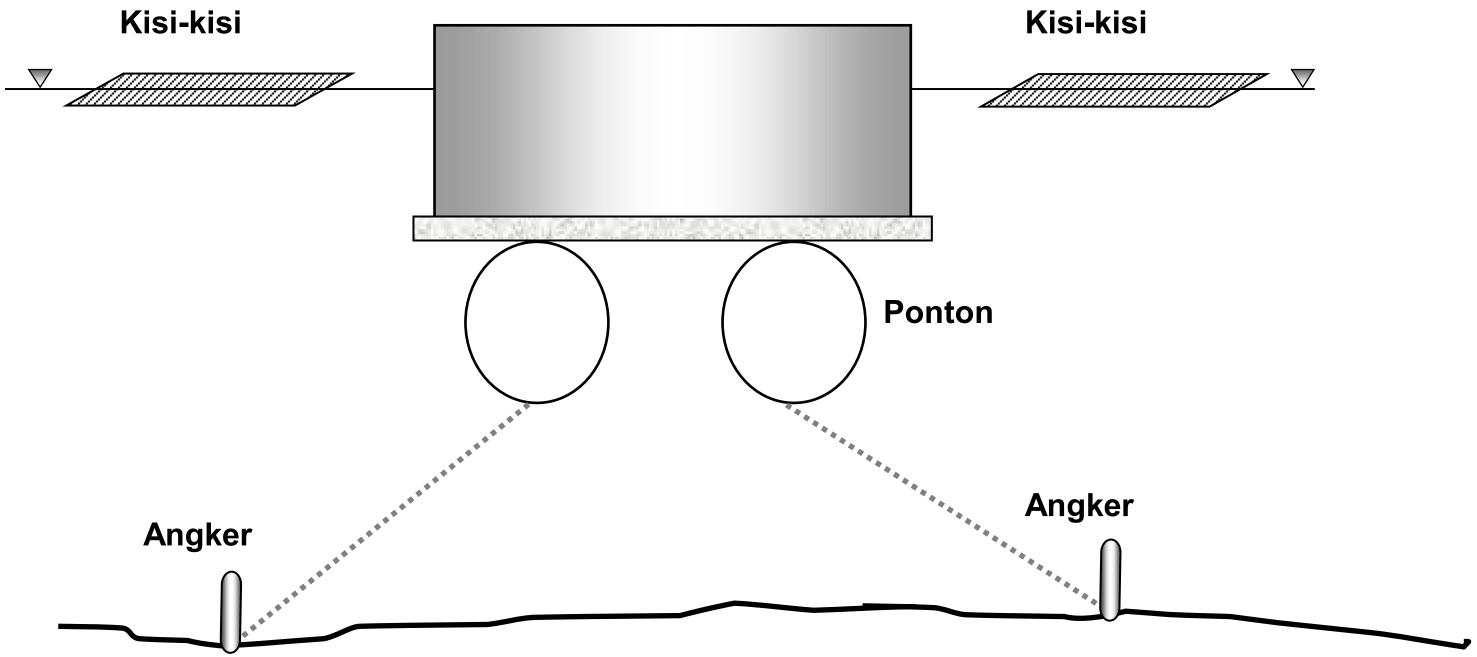
Untuk mengukur evaporasi dari muka air bebas dapat digunakan panci penguapan (*evaporation pan*). Terdapat tiga macam panci penguapan yang sering digunakan, yaitu panci penguapan klas A (*class A evaporation pan*), panci penguapan tertanam (*sunken evaporation pan*) dan panci penguapan terapung (*floating evaporation pan*). Pada prinsipnya pengukuran evaporasi dengan ketiga macam alat tersebut sama, yaitu dengan pembacaan tinggi muka air di panci pada dua saat yang berbeda sesuai dengan interval waktu pengukuran yang diinginkan. Pada setiap pengamatan umumnya juga dilakukan pengukuran temperatur air. Pan evaporasi lebih sering digunakan untuk mengukur evaporasi harian yang dinyatakan dalam mm/hari. Ilustrasi cara pemasangan panci evaporasi klas A ditunjukkan pada gambar di bawah.



Panci Evaporasi Klas A



Panci Penguapan Tertanam



Panci Penguapan Terapung

Perkiraan Evaporasi dengan Pendekatan Teoritik

Seperti telah disinggung pada uraian tentang fator-faktor yang mempengaruhi laju penguapan, pendekatan teoritik untuk perkiraan nilai penguapan memerlukan data parameter klimatologi.

Data tersebut meliputi :

- temperatur udara (*T*),
- kelembaban relatif udara atau *relative humidity* (*RH*),
- kecepatan angin pada ketinggian tertentu, yang umumnya diukur pada ketinggian 2 m di atas permukaan tanah (*U2*),
- lama penyinaran matahari atau *sunshine duration* dalam jam (*n*),
- lama penyinaran matahari maksimum pada suatu hari tertentu di lokasi pengukuran (*N*),
- radiasi matahari (*Rn*)
- dan kemungkinan data lain tergantung pada pendekatan yang digunakan untuk menurunkan rumus empiris hitungan evaporasi.

Pendekatan Hitungan Evaporasi

Setidaknya ada 3 prinsip pendekatan hitungan evaporasi, yaitu :

1. **Persamaan keseimbangan air (water balance)**

   Cara ini sangat sederhana dengan rumus berikut ini:

$$I = O \pm \Delta S$$

dengan:

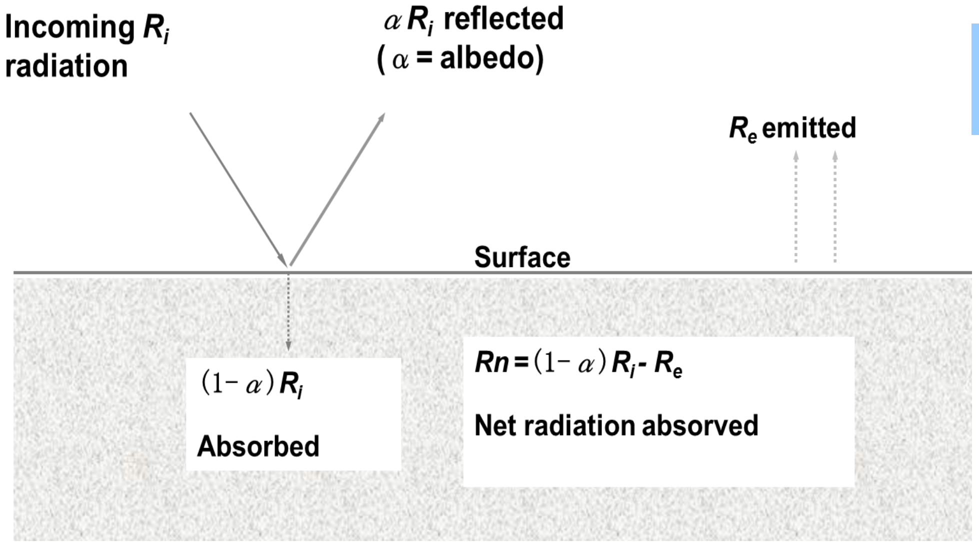
I = total inflow,

O = total outflow,

ΔS = selisih jumlah tampungan

2. Pendekatan *Energy Balance Method*

Sumber energi panas untuk proses penguapan pada permukaan air adalah perubahan panas neto (*net radiation flux*) di permukaan bumi (*Rn*). Besarnya *Rn* merupakan selisih antara serapan panas efektif di permukaan bumi dan pancaran panas ke udara (*emitted radiation*) seperti dijelaskan pada rumus dan gambar berikut ini.



$$Rn = R_i(1-\alpha) - R_e$$

3. Pendekatan Aerodynamic Method

Selain suplai energi panas, faktor lain yang mengontrol laju evaporasi adalah kemampuan untuk memindahkan uap air dari permukaan air. Proses pemindahan uap air ini akan tergantung kepada besarnya pertambahan kelembaban arah vertikal (*gradient of humidity*) dan kecepatan angin di udara dekat permukaan air. Kedua proses tersebut dapat dianalisis dengan menggunakan persamaan perpindahan massa dan momentum di udara. Penurunan rumus hitungan evaporasi dengan cara ini menghasilkan persamaan berikut (Chow, dkk., 1988):

$$E_a = B(e_{as} - e_a)$$

Ea = evaporasi dari muka air bebas selama periode pengamatan,
B = faktor empiris tergantung kepada konstanta von Karman (k), rapat massa udara (ρa), rapat massa air (ρw), kecepatan angin pada 2 m di atas permukaan (U2) dan tekanan udara ambient (p),
eas = tekanan uap jenuh di udara pada temperatur sama dengan temperatur air,
ea = tekanan uap nyata pada ketinggian pengamatan.

Rumus Hitungan Perkiraan Evapotranspirasi

Hitungan perkiraan laju evapotranspirasi, yaitu jumlah evaporasi dari permukaan tanah dan transpirasi dari tanaman juga diturunkan dengan memperhatikan faktor-faktor seperti halnya pada penurunan rumus evaporasi.

Persamaan yang umum digunakan diantaranya adalah cara ***Thornthwaite*** dan ***Penman*** .

Contoh dengan Rumus ***Thornthwaite***
Di suatu daerah yang terletak pada garis lintang 10° lintang selatan diperoleh data temperatur rerata bulanan seperti disajikan dalam tabel berikut ini.

Bulan	Jan.	Feb.	Mar.	Apr.	Mei	Juni	Juli	Agt.	Sep.	Okt.	Nov	Des.
Suhu (°C)	26,6	27,1	26,8	27,3	26,9	26,3	25,8	25,9	26,3	26,7	26,4	26,2

Hitung evapotranspirasi potensial bulanan!

Hitungan evapotranspirasi dilakukan dengan menggunakan tabel di bawah ini.
Terlebih dulu dihitung nilai I untuk seluruh bulan dan kemudian hasilnya dijumlahkan sehingga diperoleh:

$$I = \sum_{m=1}^{12} \left(\frac{T_m}{5} \right)^{1,514} = 150,11$$

Kemudian dihitung nilai a berdasar nilai I yang telah diperoleh:

$$a = 675 \cdot 10^{-9} I^{3} - 771 \cdot 10^{-7} I^{2} + 179 \cdot 10^{-4} I + 492 \cdot 10^{-3}$$

$$a = 675 \cdot 10^{-9} (150,11)^{3} - 771 \cdot 10^{-7} (150,11)^{2} + 179 \cdot 10^{-4} (150,11) + 492 \cdot 10^{-3}$$

$$a = 3,725$$

Dari nilai a dan I yang telah diperoleh dan untuk setiap nilai Tm, dihitung ET setiap bulan :

$$ET = 1{,}62\left(\frac{10 \cdot T_m}{150{,}11}\right)^{3{,}725}$$

Bulan	T_m (°C)	I	ET (cm)
Jan.	26,6	12,56	13,65
Feb.	27,1	12,92	14,63
Mar.	26,8	12,70	14,03
Apr.	27,3	13,07	15,03
Mei	26,9	12,78	14,23
Juni	26,3	12,35	13,08
Juli	25,8	11,99	12,18
Agt.	25,9	12,06	12,36
Sep.	26,3	12,35	13,08
Okt.	26,7	12,63	13,84
Nov	26,4	12,42	13,27
Des.	26,2	12,28	12,90
Jumlah		150,11	162,27

Rumus lain untuk memperkirakan nilai evapotranspirasi potensial berdasarkan gabungan pendekatan cara energy balance method dan aerodynamic method juga banyak dikembangkan. Salah satu rumus yang sering dipakai di Indonesia dan beberapa negara Asia adalah rumus Penman. Rumus Penman untuk hitungan evapotranspirasi acuan (ETo) adalah sebagai berikut:

$$ETo = c\left[W.Rn + (1 - W).f(u).(e_a - e_d)\right]$$

Dengan:

ETo = evapotranspirasi acuan (mm/hari),
W = faktor bobot temperatur,
Rn = radiasi neto ekuivalen dengan nilai evaporasi (mm/hari),
f(u) = fungsi faktor kecepatan angin,
ea-ed = selisih tekanan uap jenuh dan nyata pada temperatur udara (mbar),
c = faktor koreksi efek perubahan kondisi siang malam.

INFILTRASI

Definisi

Infiltrasi adalah proses masuknya air hujan ke dalam tanah.

Air hujan yang masuk ke dalam tanah masuk melalui proses infiltrasi, kemudian masuk ke dalam lapisan tidak kenyang air (*zone of aeration*), dan selanjutnya mengalir dalam arah lateral sebagai aliran antara (*interflow*) menuju sungai, serta mengalir secara vertikal dan dikenal dengan perkolasi (*percolation*) menuju air tanah atau bergerak ke permukaan sebagai evapotranspirasi.

Gerak air di dalam tanah melalui pori-pori tanah dipengaruhi oleh:

1. gaya gravitasi
2. gaya kapiler

Gaya gravitasi menyebabkan aliran selalu menuju ke tempat yang lebih rendah, sementara gaya kapiler menyebabkan air bergerak ke segala arah. Gaya kapiler pada tanah kering lebih besar daripada tanah basah. Selain itu gaya kapiler bekerja lebih kuat pada tanah dengan lapisan lebih halus seperti lempung daripada tanah berbutir kasar seperti pasir.

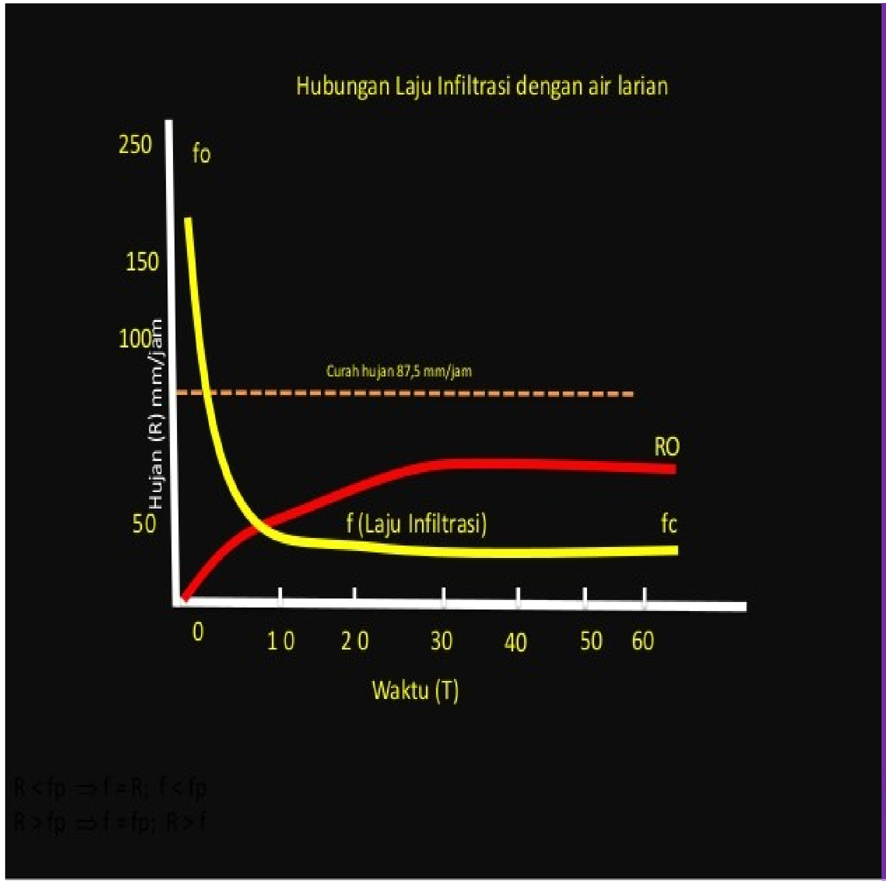
Kapasitas dan Laju Infiltrasi

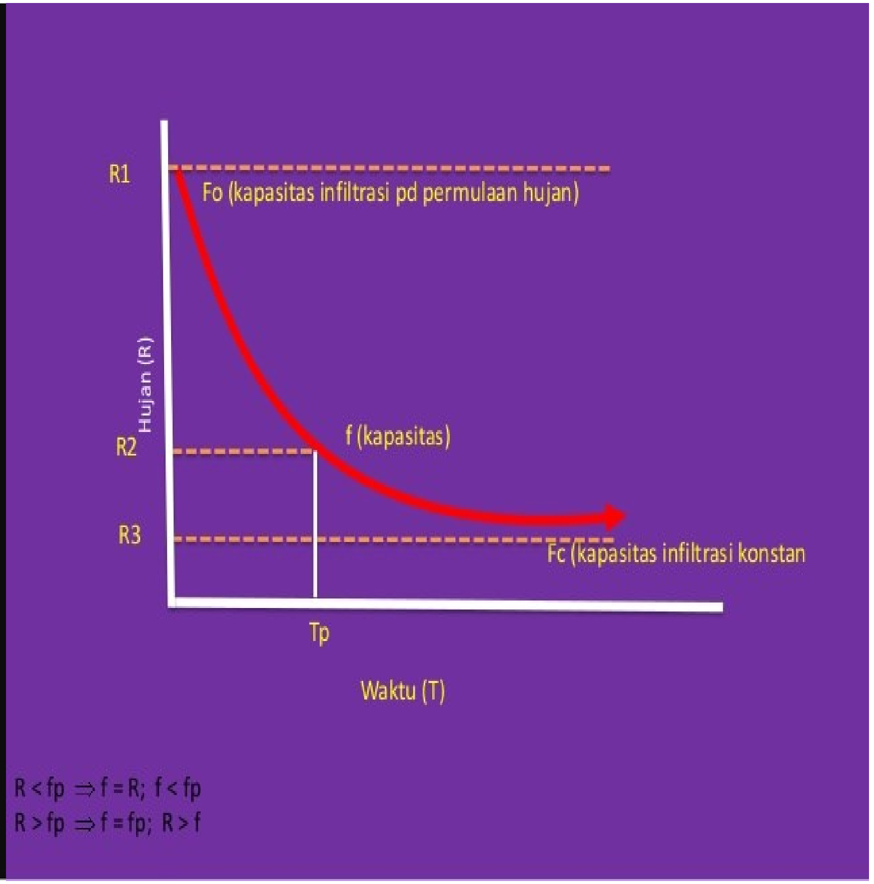
Kapasitas infiltrasi adalah laju infiltrasi maksimum untuk suatu jenis tanah tertentu.

Laju infiltrasi adalah kecepatan infiltrasi yang nilainya tergantung pada kondisi tanah dan intensitas hujan.

Kemampuan tanah menyerap air akan semakin berkurang dengan makin bertambahnya waktu. Pada tingkat awal kecepatan penyerapan air cukup tinggi dan pada tingkat waktu tertentu kecepatan penyerapan air ini akan menjadi konstan.

Kapasitas infiltrasi dapat diukur dengan menggunakan infiltrometer dan analisis hidrograf. Infiltrometer ini dibedakan menjadi dua macam yaitu infiltrometer genangan dan simulator hujan (rainfall simulators).


Hubungan Laju Infiltrasi dengan air larian
250
fo
150
100
Hujan (R) mm/jam
Curah hujan 87,5 mm/jam
RO
50
f (Laju Infiltrasi)
fc
0
1 0
2 0
30
40
50
60
Waktu (T)


R1
Fo (kapasitas infiltrasi pd permulaan hujan)
Hujan (R)
f (kapasitas)
R2
R3
Fc (kapasitas infiltrasi konstan
Tp
Waktu (T)
R < fp ⇒ f = R; f < fp
R > fp ⇒ f = fp; R > f

Faktor-faktor Infiltrasi

1. Kedalaman genangan dan tebal lapis jenuh
2. Kelembaban Tanah
3. Pemampatan oleh hujan
4. Penyumbatan oleh butir halus
5. Tanaman penutup
6. Topografi
7. Intensitas hujan

Pengukuran Infiltrasi

Dalam kaitannya dengan analisis hidrologi, informasi yang diperlukan adalah laju infiltrasi yang berubah dengan waktu.
Menurut Knapp (1978) untuk megumpulkan data infiltrasi dapat dilakukan dengan tiga cara:
(1) inflow-outflow
(2) Analisis data hujan dan hidrograf,
(3) menggunakan double ring infiltrometer.
Cara yang terakhir sering digunakan karena mudah dalam pengukuran dan alatnya mudah dipindah-pindah.

Model persamaan kurva kapasitas infiltrasi (Infiltration Capacity Curve, IC-Curve) yang dikemukakan Horton adalah sebagai berikut :

$$\boxed{f = f_c + (f_o - f_c)e^{-Kt}}$$

dimana :
f = kapasitas infiltrasi pada saat t (cm/jam)
fc = besarnya infiltrasi saat konstan (cm/jam)
fo = besarnya infiltrasi saat awal (cm/jam)
K = konstanta
t = waktu dari awal hujan
e = 2,718

Perhitungan Kapasitas Infiltrasi

Data pengukuran kapasitas infiltrasi (f) dan waktu (t) tercantum pada Tabel 1. Buatlah persamaan kurva kapasitas infiltrasi tersebut menurut model Horton?

Tabel 1 Data infiltometer (double ring)

t (jam)	0	0,25	0,50	0,75	1,00	1,25	1,50	1,75	2,00
f(cm/jam)	10,4	5,6	3,2	2,1	1,5	1,2	1,1	1,0	1,0

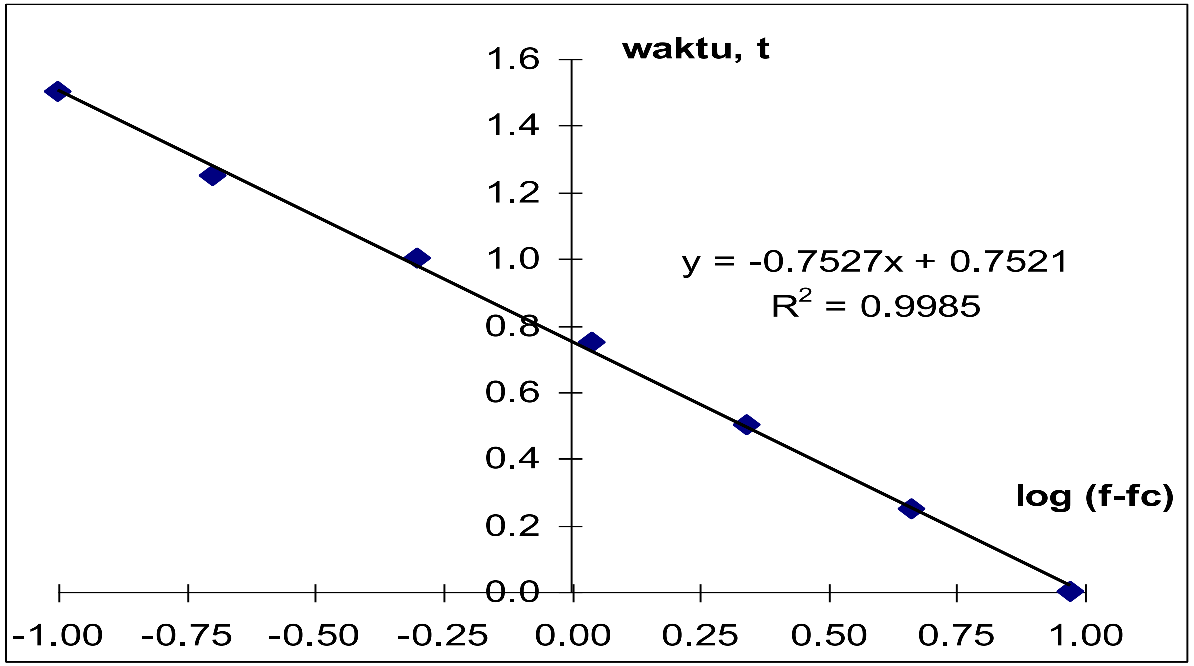
Tabel 2 Perhitungan parameter infiltrasi

Waktu (t) (jam)	kapasitas infiltrasi(f) (cm/jam)	fc	f - fc	log (f - fc)
0,00	10,4	1,0	9,4	0,973
0,25	5,6	1,0	4,6	0,663
0,50	3,2	1,0	2,2	0,342
0,75	2,1	1,0	1,1	0,041
1,00	1,5	1,0	0,5	-0,301
1,25	1,2	1,0	0,2	-0,699
1,50	1,1	1,0	0,1	-1,000
1,75	1,0	1,0	0,0	
2,00	1,0	1,0	0,0	

Persamaan liner regresi $y = m X + C$ atau $y = t$ dan $X = \log (f - fc)$

Dengan memplot hubungan t dan log (f - fc) pada kertas grafik atau menggunakan kalkulator maka diperoleh persamaan sbb.

$y = -0,7527 X + 0,7521$ (lihat Gambar)



Gambar 1 Kurva mencari gradien m

dari persamaan liner tersebut diperoleh gradien, $m = -0,7527$

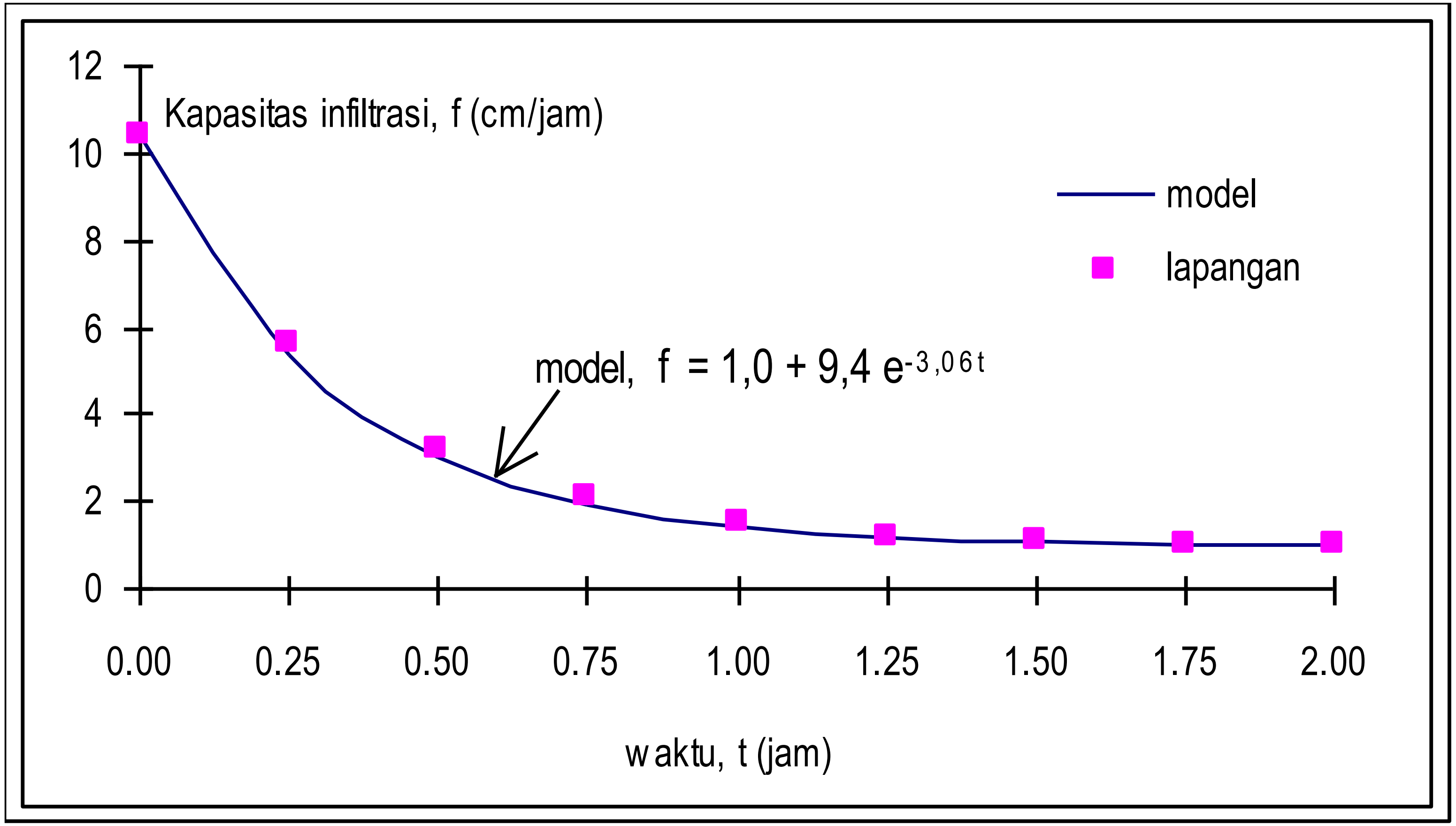
dengan menggunakan rumus $K = -1/0{,}434\ m$, maka $\mathbf{K = 3{,}06}$
dengan diketahuinya nilai pada Tabel 3.2, maka nilai

$f_c = 1.0$
$f_o = 10{,}4$
$K = 3{,}06$

maka persamaan kurva kapasitas infiltrasinya adalah

$f = f_c + (f_o - f_c)\, e^{-Kt}$ atau
$f = 1{,}0 + (10{,}4 - 1{,}0)\, e^{-3{,}06t}$ atau
$\mathbf{f = 1{,}0 + 9{,}4\, e^{-3{,}06t}}$

Gambar 2, memperlihatkan bagaimana model Horton yang digunakan dapat menduga nilai pengamatan lapangan. Ini berarti model Horton sangat tepat (fitting) dengan pengamatan lapangan.



Gambar 2 Kurva fitting persamaan model Horton

Volume Infiltrasi

Untuk menghitung jumlah infiltrasi total (Vt) selama waktu (t) maka dari persamaan Horton tersebut dilakukan integral dari persamaan Horton yang menghasilkan luasan dibawah kurva, yaitu :

$$V(t) = fc.t + \frac{(fo - fc)}{K}(1 - e^{-Kt})$$

Satuan volume total (Vt) = tinggi kolom air (mm, cm dan inchi tergantung satuan pada parameter infiltrasi yang digunakan.

Contoh :

Dari perhitungan persamaan kurva Horton di atas diperoleh, f_c = *1.0 cm/jam;* f_o = *10,4 cm/jam dan* K = 3,06. Hitung volume total infiltrasi selama 2 jam untuk areal 1 ha?.

Penyelesaian :

a. Jumlah tinggi air (2 jam) $= 1{,}0 \cdot 2 + \dfrac{(10{,}4 - 1{,}0)}{3{,}06}\left(1 - 2.718^{-3{,}06.2}\right)$

$$= 5{,}07 \text{ cm} = 0{,}0507 \text{ m}$$

b. Volume air infiltrasi pada areal 1 ha selama 2 jam adalah

$$V = 0{,}0507 \times 10^4 \text{ m}^3 = 507 \text{ m}^3$$

TERIMAKASIH

MATERI – 4

LIMPASAN

Disampaikan Oleh :
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Muhamad Komarudin S.Si., M.Si

Limpasan (Run off)

ISTN

Definisi

Limpasan permukaan adalah aliran air yang mengalir di atas permukaan karena penuhnya kapasitas infiltrasi tanah. Limpasan ini terjadi apabila intensitas hujan yang jatuh di suatu DAS melebihi kapasitas infiltrasi, setelah laju infiltrasi terpenuhi maka air akan mengisi cekungan-cekungan pada permukaan tanah. Setelah cekungan-cekungan tersebut penuh, selanjutnya air akan mengalir (melimpas) diatas permukaan tanah

Faktor Yang Mempengaruhi Limpasan

Elemen Meteorologi

- Jenis Presipitasi (Salju dan Hujan)
- Intensitas Curah Hujan
- Lamanya Curah Hujan
- Distribusi curah hujan dalam D.A.S.
- Curah hujan sebelumnya dan lembab tanah (Soil Moisture).

2. Elemen D.A.S.
   - Kondisi permukaan tanah (Landuse)
   - Kondisi Topografi dan bentuk D.A.S.
   - Jenis Tanah menentukan kapasitas infiltrasi
3. Faktor manusia (Seyhan, 1977).

Analisa Limpasan Curah Hujan

Menghitung limpasan permukaan (run off) pada suatu area lahan penting untuk maksud perencanaan penggunaan lahan. Dari perhitungan pendugaan run off itu dapat dibuat perencanaan untuk berbagai hal, salah satunya adalah upaya apa yang dapat dilakukan dalam rangka mengendalikan run off dan erosi tanah. Selain itu, para perencana dapat merencanakan pembuatan waduk, palung atau hanya cek dam atau embung dalam rangka melakukan konservasi air.

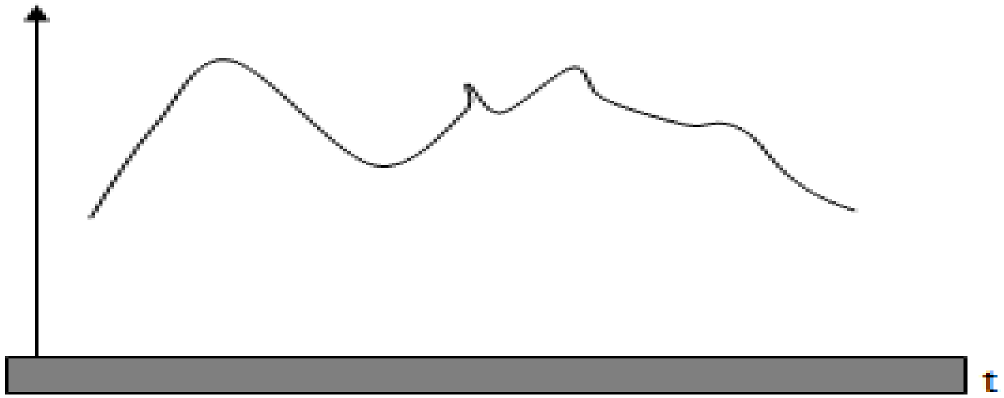
Dengan menggunakan rumus Rasional, pendugaan debit air limpasan dapat dilakukan dengan mudah. Debit air limpasan adalah volume air hujan per satuan waktu yang tidak mengalami infiltrasi sehingga harus dialirkan melalui saluran drainase. Debit air limpasan terdiri dari tiga komponen yaitu Koefisien Run Off (C), Data Intensitas Curah Hujan (I), dan Catchment Area (Aca).

Koefisien yang digunakan untuk menunjukkan berapa bagian dari air hujan yang harus dialirkan melalui saluran drainase karena tidak mengalami penyerapan ke dalam tanah (infiltrasi).

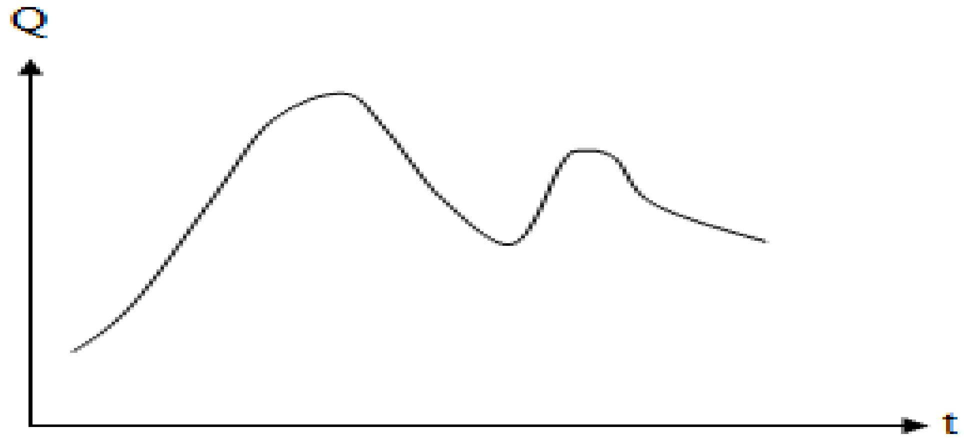
Perkiraan Debit Banjir

Hubungan antara debit dan tinggi muka air dapat dihitung dengan menggunakan stage hydrograph curve. Hidrograf adalah suatu diagram yang menggambarkan variasi debit sungai atau tinggi muka air menurut waktu Hidrograf menunjukkan tanggapan menyeluruh DAS terhadap masukan tertentu. Sesuai dengan sifat dan perilaku DAS yang bersangkutan, hidrograf aliran selalu berubah sesuai dengan besaran dan waktu terjadinya masukan. Bentuk hidrograf banjir sangat dipengaruhi oleh bentuk DAS. Jika bentuk DAS membesar di tengah maka bentuk hidrografnya adalah debit puncak berlangsung dalam waktu yang cepat. Jika berbentuk membesar di hulu maka debit puncak akan dicapai dalam waktu yang relatif lama, sedangkan jika berbentuk mengecil ditengah dan membesar dibagian hulu dan hilir maka bentuk hidrografnya mempunyai puncak dua buah. Jika DAS mempunyai bentuk panjang maka bentuk hidrografnya relatif simetris.

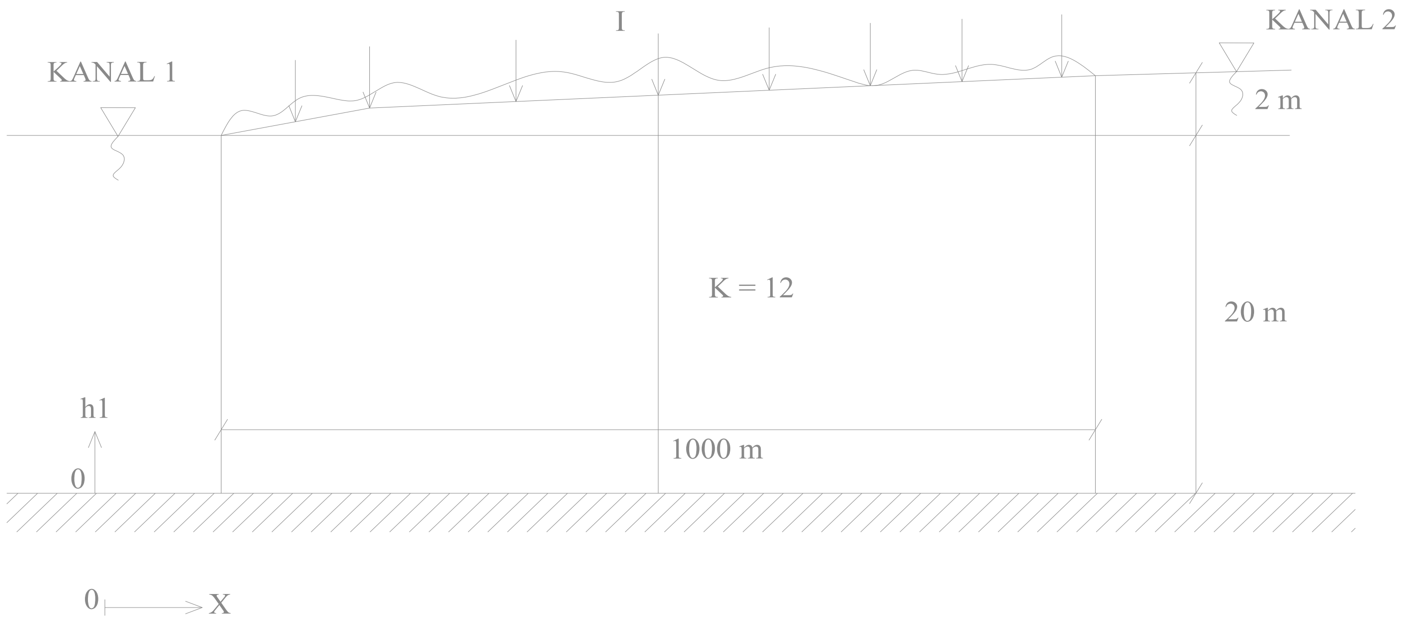
Hubungan waktu (t) dan tinggi Muka Air (H) disebut hidrograph air (Stage hydrograph)



Hubungan waktu (t) dan Debit (Q) disebut hidrograph Debit (discharge hydrograph)



Perhitungan Air Tanah



Hitung :

Aliran air tanah antara 2 kanal yang berjarak 1000 meter
Koef. Permeabilitas tanah = 12 m/hari
Benda tinggi M.A. pada kedua kanal = 2m
Kedalaman akifer = 20 m dibawah permukaan air kanal 1
Hujan tahunan = 1,2 m/tahun
Diasumsikan infiltrasi = 60% dari hujan

Jawaban :

Dengan mengambil titik 0 sebagai titik pusat kordinat (seperti pada gambar) ;
Maka :

- Kondisi batas (boundary conditions)

  $x_1 = 0 \longrightarrow h_1 = 20$ meter

  $x_2 = 1000$ m $\longrightarrow h_2 = 22$ meter

- $I = 0{,}60 \times 1{,}2$ meter / tahun

  $= 0.72$ m/tahun

  $= \dfrac{0.72}{365}$ m / hari

$$\frac{d^2(h^2)}{dx^2} = \frac{2.I}{k_x}$$

$$\frac{d^2(h^2)}{dx} = \frac{2.I}{k_x} . X + C_1$$

$$h^2 = -\frac{I}{k_x} . X^2 + C_1 . X + C_2$$

$x_1 =$ —— $h_1 = 20$ meter sehingga

$400 = -0 + 0 + C_2$

$C_2 = 400$

$X_2 = 1000 \qquad h_2 = 22$ meter sehingga

$$484 = -\frac{0.72}{365} . 10^6 + C_1 . 1000 + 400$$

$C_1 = 0.084 + 0.164 = 0.248$

$$h = \sqrt{\left[-\frac{I}{k_x} \cdot x^2 + 0{,}248\ x + 400 \right]}$$

$$= \mu^{1/2}\ , \text{ dengan } \mu = -\frac{I}{k_x} \cdot x^2 + 0{,}248\ x + 400$$

$$\frac{dh}{dx} = \frac{dh}{du}\frac{du}{dx} = \frac{1}{2\ U^{1/2}} \cdot \left[-\frac{I}{k_x} \cdot 2\ x + 0{,}248 \right] =$$

$$= \frac{I}{2.h} - \left[\frac{I}{k_x} .2\ x + 0.248 \right]$$

- Dari pers. (3) : $q_x = -\,k_x \cdot h \cdot \dfrac{dh}{dx}$ $\quad hitung\ q_x$

$x = 0 \quad \mu = 400;\ h = \sqrt{400} = 20;$

$$\frac{dh}{dx} = \frac{1}{2 \text{ x } 20}\ (-0 + 0.248)$$

$$q_x = -\,12 \text{ x } 20 \text{ x } \frac{1}{2 \text{ x } 20} \cdot (0.248) = -\,1.49\ m^2\ /\ hari\ /\ m$$

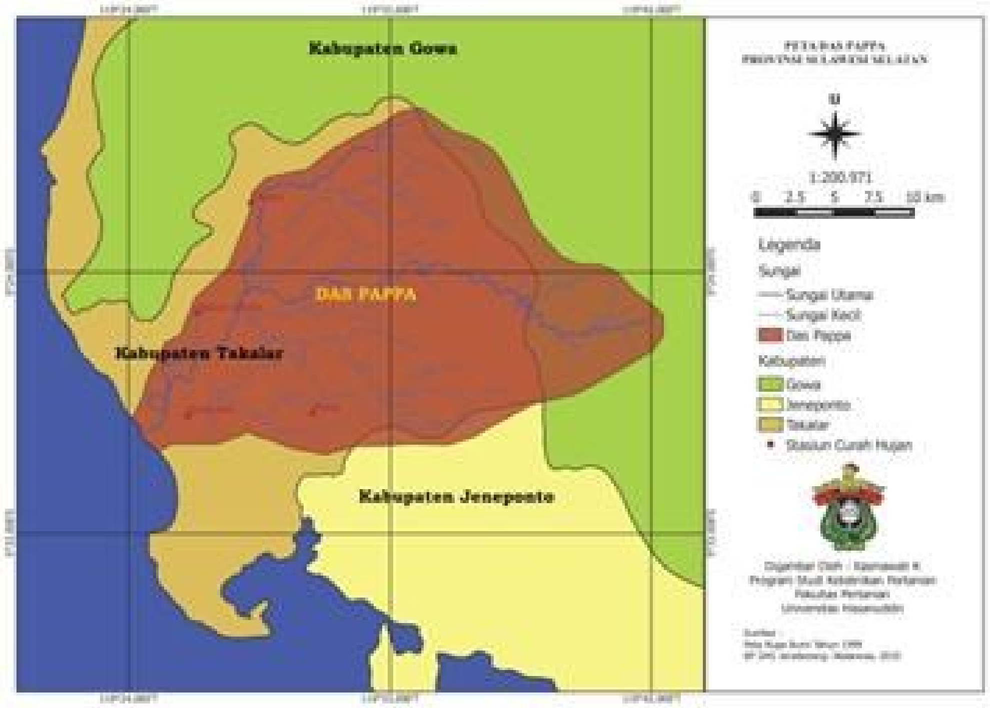
$$x = 1000 \qquad q_x = -\,12.22 \left[\frac{1}{2.22} \left[-\frac{0.72 \text{ x } 2 \text{ x } 1000}{365 \text{ x } 12} + 0.248 \right] \right]$$

$(- 0.328 + 0.248) = 0.48\ m^3$ / hari / m

ISTN

1. Daerah Aliran Sungai Pappa

Sungai Pappa berada di bagian barat Sulawesi Selatan, secara administrasi terletak di Kabupaten Takalar, dengan panjang sungai 57 km dan luas daerah pengaliran 389 km2.



Bagian hulu DAS Pappa berada pada sungai Pamukkulu dengan ketinggian ±750 dpl yang terletak di Kab. Gowa.

2. Analisis Curah Hujan Rencana

Curah hujan harian maksimum rerata daerah diperoleh dari 5 stasiun yang berpengaruh pada DAS Pappa, berdasarkan metode poligon Thiessen, dengan luas dan koefisien Thiessen disajikan pada Tabel 1.

Tabel 1.Koefisien *Thiessen*stasiun hujan DAS Pappa

No	Stasiun Hujan	Luas (km^2)	KT
1	Pamukkulu	214	0.55
2	Paleko	31	0.08
3	Malolo	66	0,17
4	Bontocinde	55	0,14
5	Pabentengan	23	0.06

Curah hujan maksimum rerata daerah 342,15 mm dan curah hujan rerata minimum 53,75 mm, hal ini disebabkan oleh perbedaan intensitas hujan yang terjadi.
Sedangkan analisis frekwensi curah hujan DAS Pappa, menggunakan distribusi Gumbel dan Log Pearson III. Hasil uji kesesuain distribusi disajikan pada Tabel 2.

Tabel 2. Kesesuaian distribusi frekwensi dengan uji Chi-Kuadrat

No.	Metode Distribusi	Peluang (%)
1	Gumbel	1,66
2	Log Pearson III	8,63

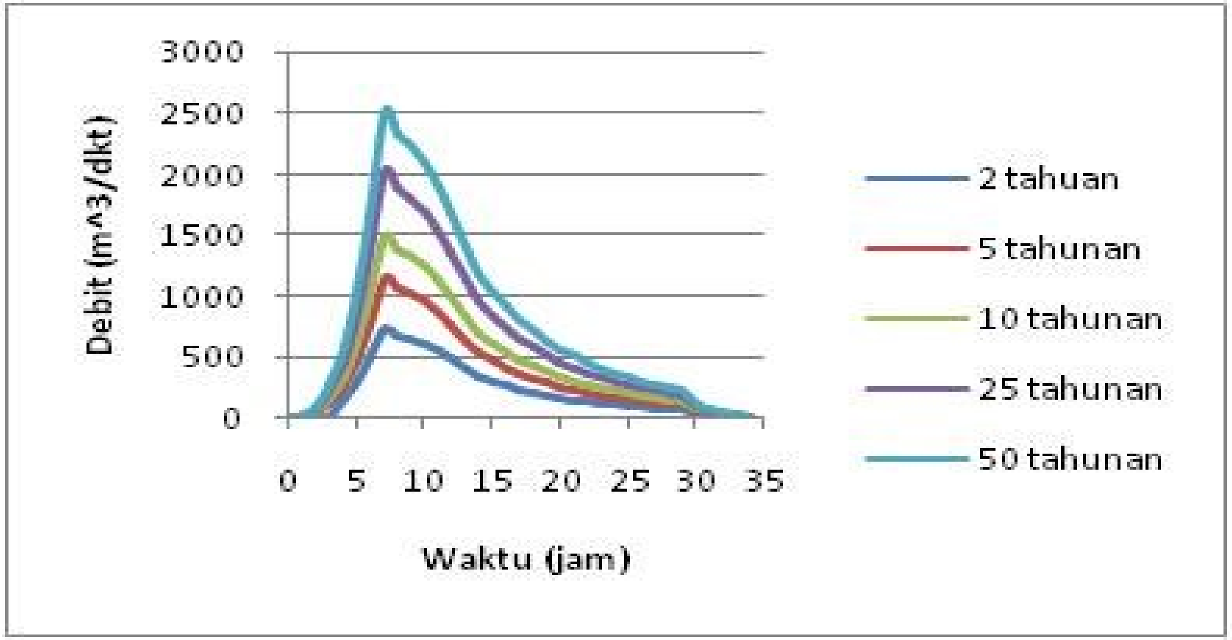
Hasil uji chi-kuadrat pada perhitungan distribusi Gumbel dan Log Pearson Tipe III, maka distribusi curah hujan yang paling sesuai untuk DAS Pappa adalah distribusi Log Pearson III. Sehingga curah hujan rencana berdasarkan distribusi Log Pearson III disajikan pada Tabel 3.

Tabel 3. Curah hujan rencana DAS Pappa

No.	Periode ulang (tahun)	Curah hujan rencana (mm/hari)
1	2	98
2	5	154
3	10	201
4	25	273
5	50	339

3. Debit Banjir Maksimum

Hidrograf Satuan Sintetik Nakayasu merupakan metode yang didasarkan pada pola distribusi hujan dan hujan efektif yang jatuh merata dalam selang waktu 6 jam, menurut rasio intensitasnya.



Gambar Debit banjir maksimum DAS Pappa

Berdasarkan grafik Gambar 3., diketahui bahwa waktu yang dibutuhkan dari permulaan hujan hingga diperoleh banjir maksimum adalah 6,7 jam. Besarnya debit banjir maksimum ditunjukan oleh nilai debit pada titik puncak kurva pada hidrograf satuan, yang disajikan pada Tabel 4.

Tabel 4. Debit banjir maksimum rencana

No.	**Periode ulang (tahun)**	**Debit banjir maksimum (m^3/dkt)**
1	2	726,79
2	5	1142,13
3	10	1490,55
4	25	2024.65
5	50	2514,14

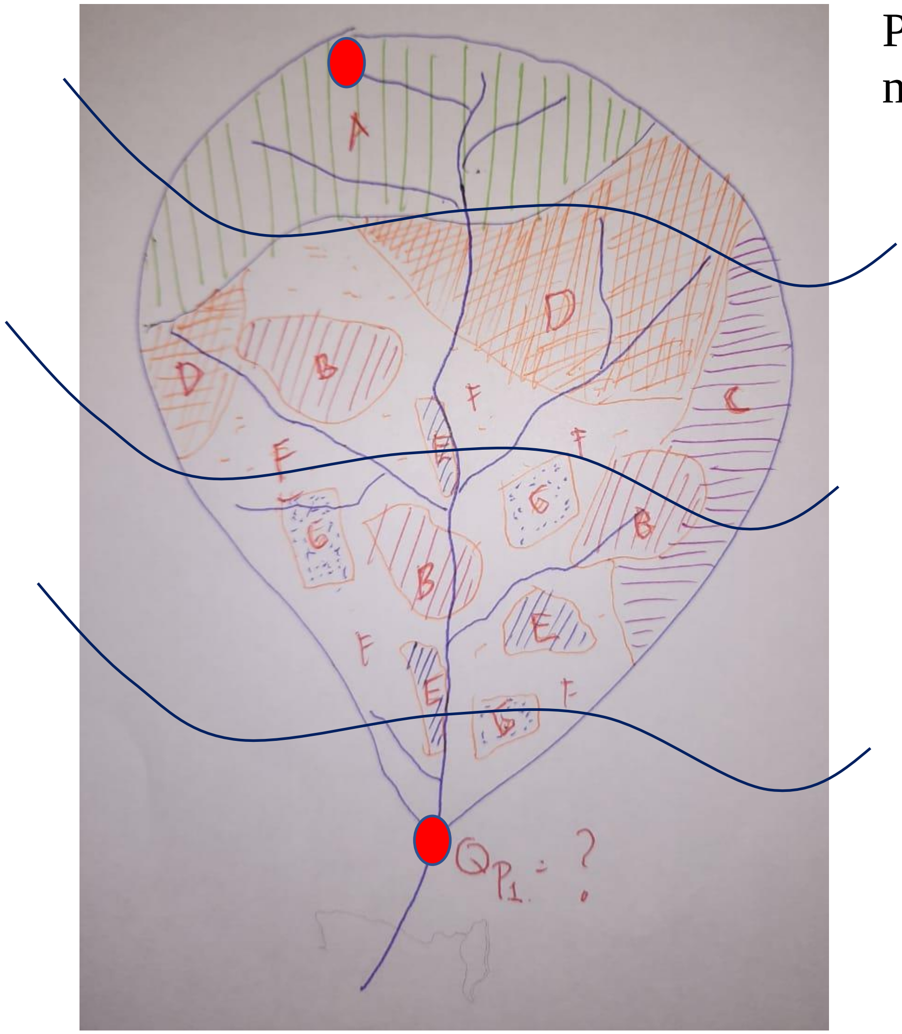
Debit banjir maksimum dapat digunakan untuk perencanaan bangunan air. Pemilihan periode ulang dipengaruhi oleh beberapa faktor, diantaranya jenis bangunan air yang direncanakan, karakteristik sungai, jumlah penduduk, periode kejadian bencana dan lain sebagainya.

Curah hujan harian rerata maksimum DAS Pappa 342,15 mm dan minimum 53,75 mm. Distribusi CH yang sesuai untuk menghitung CH rencana adalah Log Person III. Debit banjir maksimum untuk periode ulang 5, 10 dan 50 tahunan masing-masing 1142,13, 1490,55 dan 2514,14 m3/dtk, dengan waktu (Tp = time peak) 6,7 jam.

TERIMAKASIH

CONTOH PERHITUNGAN

Debit Puncak Rencana dengan Metode Rasional



Perhitungan besarnya debit banjir rencana dengan metode rasional menggunakan rumus sebagai berikut :

$$Q_t = 0{,}278 \times C \times I_T \times A$$

Keterangan :

Q_t = debit banjir (m^3/dtk)

C = koefisien pengaliran

I_T = intensitas curah hujan dengan periode ulang T tahun (mm/jam)

A = luas areal (km^2)

Diketahui :

- Panjang lintasan air dari titik terjauh sampai titik yang ditinjau 67 km
- Panjang sungai utama (L) = 38 km
- Luas Daerah aliran (A) = 544,32 km2
- Elevasi Hulu = 600 m = 0,6 km
- Elevasi Hilir = 30 m = 0,03 km
- Elevasi titik terjauh = 700 m = 0,7 km
- Tebal Hujan Harian Maksimum Periode Ulang 2 Tahun : 64.8

L = L1 + L2
= 38 + 29 = 67

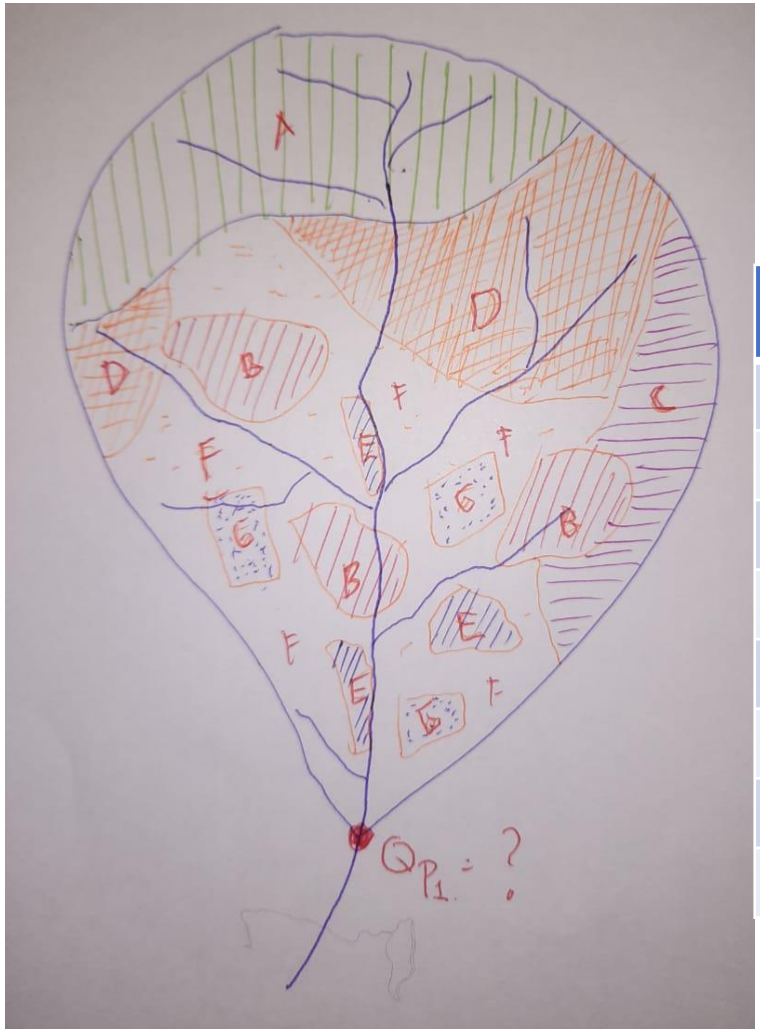
Kemiringan S1 = (0,7-0,6)/29=0,003

Kemiringan S2 = (0,6-0,03)/38=0,015

Kemiringan Rerata =(0,003+0,015)/2
= 0,009

CONTOH PERHITUNGAN

Debit Puncak Rencana dengan Metode Rasional



$$Q_t = 0{,}278 \text{ x } C \text{ x } I_T \text{ x } A$$

Tabel Koefisien Aliran (C)

Penggunaan Lahan	Luas (Km2)	Koeffisien Aliran C
A. Hutan	12,923	0,4
B. Pemukiman	2,232	0,3
C. Belukar	78,926	0,07
D. Perkebunan	326,320	0,4
E. Rawa	2,123	0,05
F. Tanah Terbuka	1,687	0,5
G. Tambak	0,109	0,05
Total Luas	544,32	

Perhitungan Koefisien Aliran (C)

$$\text{Hitung nilai } C_{100\%} = \frac{\sum_{i=1}^{n} C_i . A_i}{\sum_{i=1}^{n} A_i}$$

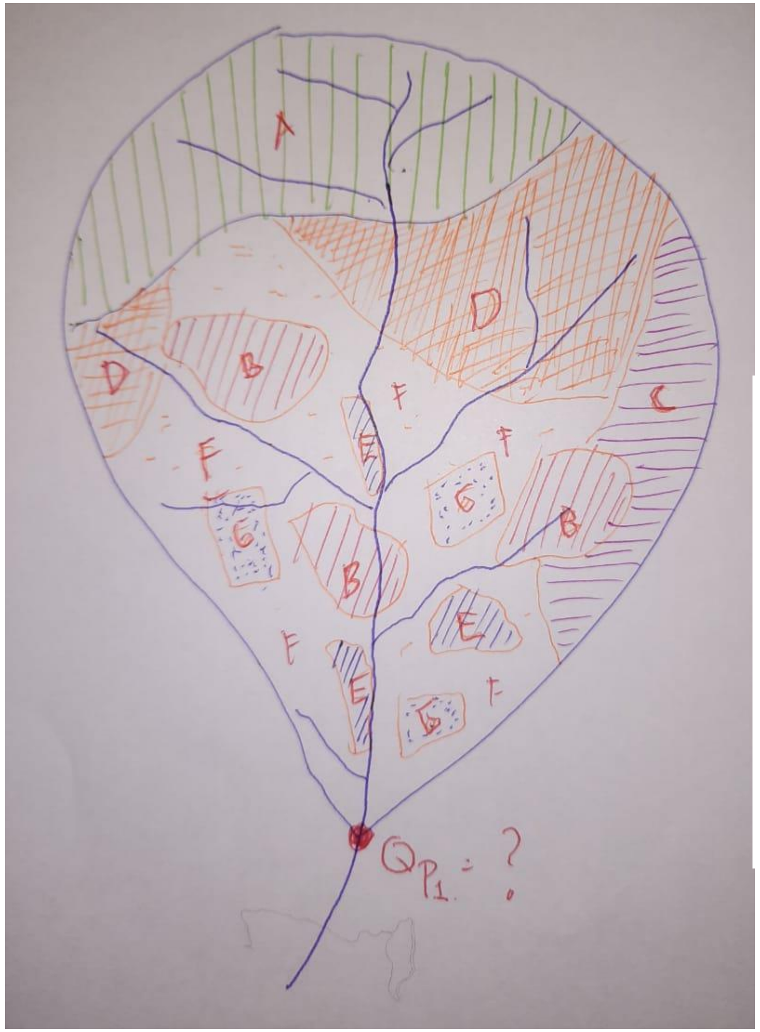
$$= \frac{C_1.A_1 + C_2.A_2 + C_3.A_3 + C_4.A_4 + C_5.A_5 + C_6.A_6 + C_7.A_7}{A}$$

$$= \frac{(0.40 \times 2{,}232) + (0.30 \times 12{,}923) + (0.07 \times 78{,}926) + (0.40 \times 26{,}32) + (0.05 \times 2{,}123) + (0.50 \times 1{,}687) + (0{,}05 \times 0{,}109)}{544{,}32}$$

$$= 0{,}26$$

CONTOH PERHITUNGAN

Debit Puncak Rencana dengan Metode Rasional



$$Q_t = 0{,}278 \text{ x } C \text{ x } I_T \text{ x } A$$

Intensitas Hujan dalam mm/jam , menurut Mononobe :

$$I_2 = \left[\frac{R}{24}\right] \text{x} \left[\frac{24}{tc}\right]^{2/3}$$

Untuk perhitungan pada periode 2 tahun :

❖ $I_2 = \left[\frac{R}{24}\right] \text{x} \left[\frac{24}{tc}\right]^{2/3}$

$= \left[\frac{64{,}38}{24}\right] \text{x} \left[\frac{24}{10{,}26}\right]^{2/3}$

$= 3{,}36$ mm/jam

Perhitungan Tc

Dik:

L	= L1+L2	
	= 38+29	= 67 km
Kemiringan (S1)	= (0,7-0,6)/29	= 0,003
Kemiringan (S2)	= (0,6-0,03)/38	= 0,015
S rerata	= (0,003+ 0,015)/2	= 0,009

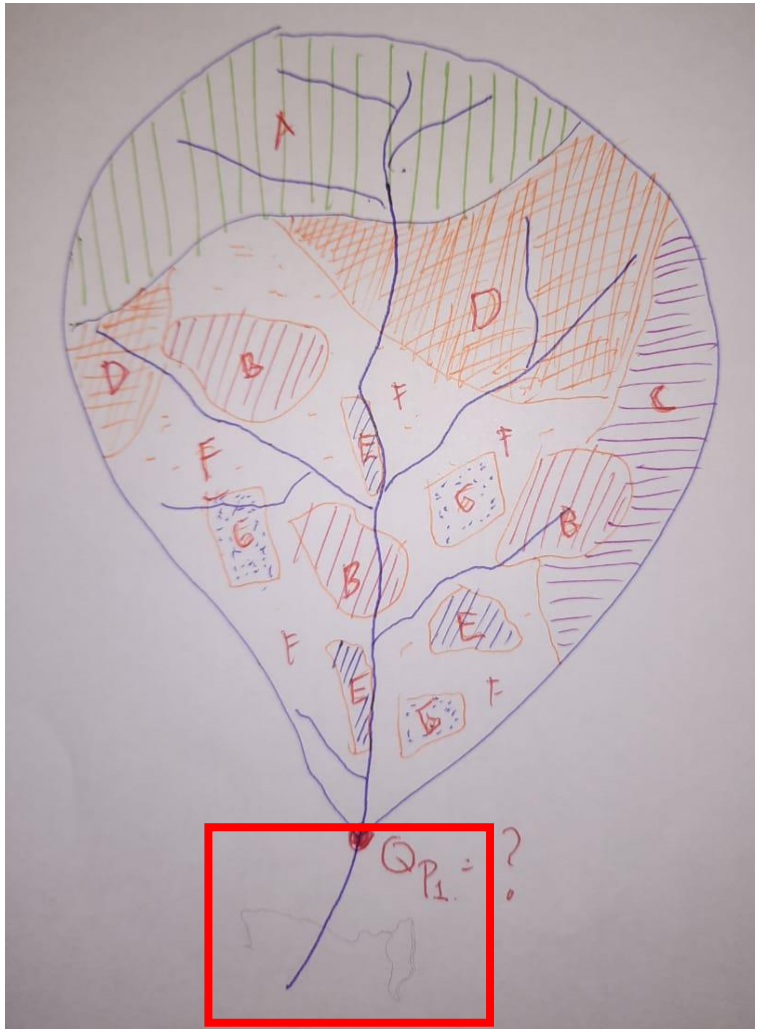
❖ $tc = \left(\frac{0{,}87 \,.\, L^2}{1000 \,.\, S}\right)^{0{,}385}$

$= \left(\frac{0{,}87 \,.\, 67^2}{1000 \,.\, 0{,}009}\right)^{0{,}385}$

$= 10{,}26$ jam

CONTOH PERHITUNGAN

Debit Puncak Rencana dengan Metode Rasional



$$Q_t = 0{,}278 \times C \times I_T \times A$$

- Untuk T = 2 tahun

$$\begin{aligned} Q_2 &= 0{,}278 \times C \times I_T \times A \\ &= 0{,}278 \times 0{,}26 \times 3{,}36 \times 544{,}32 \\ &= 132{,}43 \ \text{m}^3/\text{dtk} \end{aligned}$$

Jadi debit banjir rencana periode 2 tahun adalah 132,43 m^3/dtk
Untuk perhitungan keseluruhan dapat dilihat pada tabel di bawah ini :

periode T (tahun)	Rmax (mm)	tc (jam)	I (mm/jam)	C	A(km)	Q (m/detik)
2	64,38	10,26	3,36	0,26	544,32	132,43
5	75,76		3,74			147,44
10	83,29		3,98			156,96
25	92,81		4,28			168,57
50	99,87		4,49			176,93
100	106,88		4,69			185,03

METODE RASIONAL

Pemodelan Debit Puncak Aliran

Muhamad Komarudin

METODE RASIONAL

Implementasi Metode Rasional

- Menggambarkan hubungan antara debit limpasan dengan besar curah hujan
- Menentukan debit puncak banjir / limpasan kala ulang hujan tertentu bagi saluran-saluran dengan daerah aliran kecil, kira-kira 5000 ha.
- Curah hujan diasumsikan merata di daerah tangkapannya

Metode rasional ini dapat dinyatakan secara aljabar dengan persamaan sebagai berikut (Subarkah, 1980) :

$$Q_p = 0,00278 \text{ C.I.A}$$

Keterangan:

Qp = debit puncak banjir (m3/s);
C = koefisien limpasan;
I = intensitas hujan selama waktu konsentrasi (mm/jam)
A = luas daerah aliran (Ha).

Menghitung Koefisien Limpasan (C)

Koefisien *runoff* untuk drainase perkotaan sangat dipengaruhi oleh daerah kedap air dan dirumuskan seperti berikut :

$$C = 0{,}9\, I_m + (1 - I_m) C_p$$

Keterangan:

C_p = koefisien limpasan *runoff* untuk daerah tidak kedap air;

I_m adalah rasio kedap air.

$$I_m = \frac{A_{KEDAPAIR}}{A_{TOTAL}}$$

Tata guna lahan	Karakteristik	C	Im (%)	Keterangan
Pusat perbelanjaan dan perkantoran		0,90	100	
Industri	Bangunan penuh	0,80	80	Berkurang untuk bangunan tidak penuh
Pemukiman (kepadatan menengah – tinggi)	20 rmh/ha 30 rmh/ha 40 rmh/ha 60 rmh/ha	0,48 0,55 0,65 0,75	30 40 60 75	Bandingkan daerah kedap air dengan daearah lain
Pemukiman (kepadatan rendah)	10 rmh/ha	0,40	< 20	CN =85 (*Curve Number*)
Taman	Daerah datar	0,30	0	
Pedesaan	Tanah berpasir Tanah berat (heavy soil) Daerah irigasi		0 0 0	C = 0,20; CN = 60 C = 0,35; CN = 75 C = 0,50; CN = 85

Sumber: *USDepartment of Agriculture, Natural Resources Conservation Service, June 1986*
CATATAN : CN : *Curve Number* dari SCS

Menghitung Koefisien Limpasan (C)

Untuk C komposit dapat dihitung dengan persamaan seperti berikut :

$$C_k = \frac{C_1.A_1 + C_2.A_2 + \ldots + C_n.A_n}{A_{TOTAL}}$$

Menghitung Koefisien Limpasan (C)

Koefisien Limpasan (C) berdasarkan penggunaan lahan dapat pula ditentukan dengan menggunakan Tabel di bawah ini :

Karakteristik tanah	Tata guna lahan	Koefisien Limpasan (C)
Campuran pasir dan/ atau campuran kerikil	Pertanian	0,20
	Padang rumput	0,15
	Hutan	0,10
Geluh dan sejenisnya	Pertanian	0,40
	Padang rumput	0,35
	Hutan	0,30
Lempung dan sejenisnya	Pertanian	0,50
	Padang rumput	0,45
	Hutan	0,40

Jenis Daerah	Koefisien Aliran	Kondisi Permukaan	Koefisien Aliran
Daerah Perdagangan		Jalan Aspal	
Kota	0,70-0,95	Aspal dan beton	0,75-0,95
Sekitar kita	0,50-0,70	Batu bata dan batako	0,70-0,85
Daerah Pemukiman		Atap Rumah	0,70-0,95
Satu rumah	0,30-0,50	Halaman berumput, tanah pasir	
Banyak Rumah,terpisah	0,40-0,60	Datar, 2%	0,05-0,10
Banyak Rumah, rapat	0,60-0,75	Rata-rata,2-7 %	0,10-0,15
Pemukiman, pinggiran Kota	0,25-0,40	Curam, 7 % atau lebih	0,15-0,20
Apartemen	0,50-0,70		
Daerah Industri		Halaman berumput,tanah pasir padat	
Ringan	0,50-0,80	Datar, 2 %	0,13-0,17
Padat	0,60-0,90	Rata-Rata, 2-7 %	0,18-0,22
Lapangan, kuburan dan sejenisnya	0,10-0,25	Curam, 7 % atau lebih	0,25-0,35
Halaman, jalan kereta api dan sejenisnya	0,20-0,35		
Lahan tidak terpelihara	0,10-0,30		

Menghitung Waktu Konsentrasi (tc)

1) Cara menghitung t_c, Kirpich (1940)

$$t_c = 0{,}0195\ L^{0{,}77}\ S^{-0{,}385}$$

Keterangan:

t_c = waktu dalam menit;

L = panjang lereng dalam; atau Panjang lintasan air dari titik terjauh sampai titik tinjau (m')

S = kemiringan lereng rata-rata (m'/m').

2) Cara menghitung tc,Izzard (1994)

$$t_C = \frac{526,4 * K * L}{i^{2/3}} \quad \text{menit} \rightarrow \text{untuk } i * L < 3871$$

keterangan :

L adalah panjang aliran di lahan (m)

i adalah intensitas hujan (mm/jam)

$$K = \frac{2,756 * 10^{-4} * i + C_r}{S^{2/3}}$$

Keterangan:

L adalah panjang aliran di lahan (*overland flow distance*), dalam m;

i adalah intensitas hujan,dalam mm/jam;

S adalah slope (m/m);

Cr adalah koefisien runoff / aliran (Tabel)

Aspal Halus	**0,0070**
Aspal dan Perkerasan Pasir	**0,0075**
Atap	**0,0082**
Beton	**0,0120**
Aspal dan Perkerasan Krikil	**0,0170**
Rumput	**0,0460**
Alang-Alang	**0,0600**

3) Cara menghitung tc, Kerby (1959)

$$t_C = 1{,}44 * \left(L * n * S^{-0{,}5}\right)^{0{,}467} \text{ menit} \rightarrow \text{untuk L} < 365 \text{ m}$$

Keterangan:

L adalah panjang aliran (m);

S adalah slope (m/m);

n adalah koefisien kekasaran. (Tabel)

Tabel Besarnya koefisien kekasaran (n)

Sumber: Chin, 2000

Paving halus	0,02
Tanah terbuka	0,1
Rumput gersang / tanah terbuka	0,3
Rumput sedang	0,4
Hutan meranggas	0,6
Rumput lebat	0,8

4) Cara menghitung tc, FAA

Daerah perkotaan dengan panjang aliran antara 60 m – 100 m dirumuskan :

$$t_C = \frac{0{,}552 * \left[1{,}8 * (1{,}1 - C) * L^{0{,}5}\right]}{S^{1/3}} \text{ menit.}$$

Keterangan:

C	adalah koefisien runoff;
L	adalah panjang aliran di lahan (m);
S	adalah kemiringan lahan (%).

5) Waktu Konsentrasi di saluran:

$$t_d = \frac{L}{60 * V} \text{ menit}$$

Keterangan:

L	adalah panjang saluran (m);
V	adalah kecepatan aliran rata-rata (m/s).

Menghitung Intensitas Hujan (I)

Rumus Mononobe :

- Intensitas curah hujan adalah besarnya jumlah hujan yang turun yang dinyatakan dalam tinggi curah hujan atau volume hujan tiap satuan waktu.
- Besarnya intensitas hujan berbeda-beda, tergantung dari lamanya curah hujan dan frekuensi kejadiannya

$$I = \frac{R24}{24}\left(\frac{24}{t}\right)^{2/3}$$

Dimana :

I : Intensitas hujan (mm/jam)

R24 : curah hujan harian maksimum untuk kala ulang T (mm/jam)

t : Lama hujan (jam)

Contoh perhitungan debit puncak aliran (Qp)

Suatu wilayah Kawasan industri memiliki karakteristik daerah tangkapan sebagai berikut :

- Luas Tangkapan (A) : 44 HA
- Panjang Aliran (L) : 1050 m
- Rata-rata Kemiringan (S) : 0,0021 m / m
- Penguunaan Lahan : Kawasan Industri padat

  Dari table diperoleh nilai Koefisien Limpasan (C) : 0,75
- Curah Hujan periode ulang 10 tahun : 171,1 mm
- Waktu konsentrasi menurut metode Kirpich :

  $tc = 0{,}0195\ L^{0{,}77}\ S^{-0{,}385}$ (menit)

  = 0,0195*1050^0,77*0,0021^-0,386

  = 44,67 menit = 0,75 jam

Menghitung Intensitas Hujan (I)

Diketahui curah hujan rencana (R) sebesar 123.160 mm pada kala ulang 2 tahun, dengan lama hujan (t) adalah 1 jam. maka perhitungan Intensitas adalah sebagai berikut:

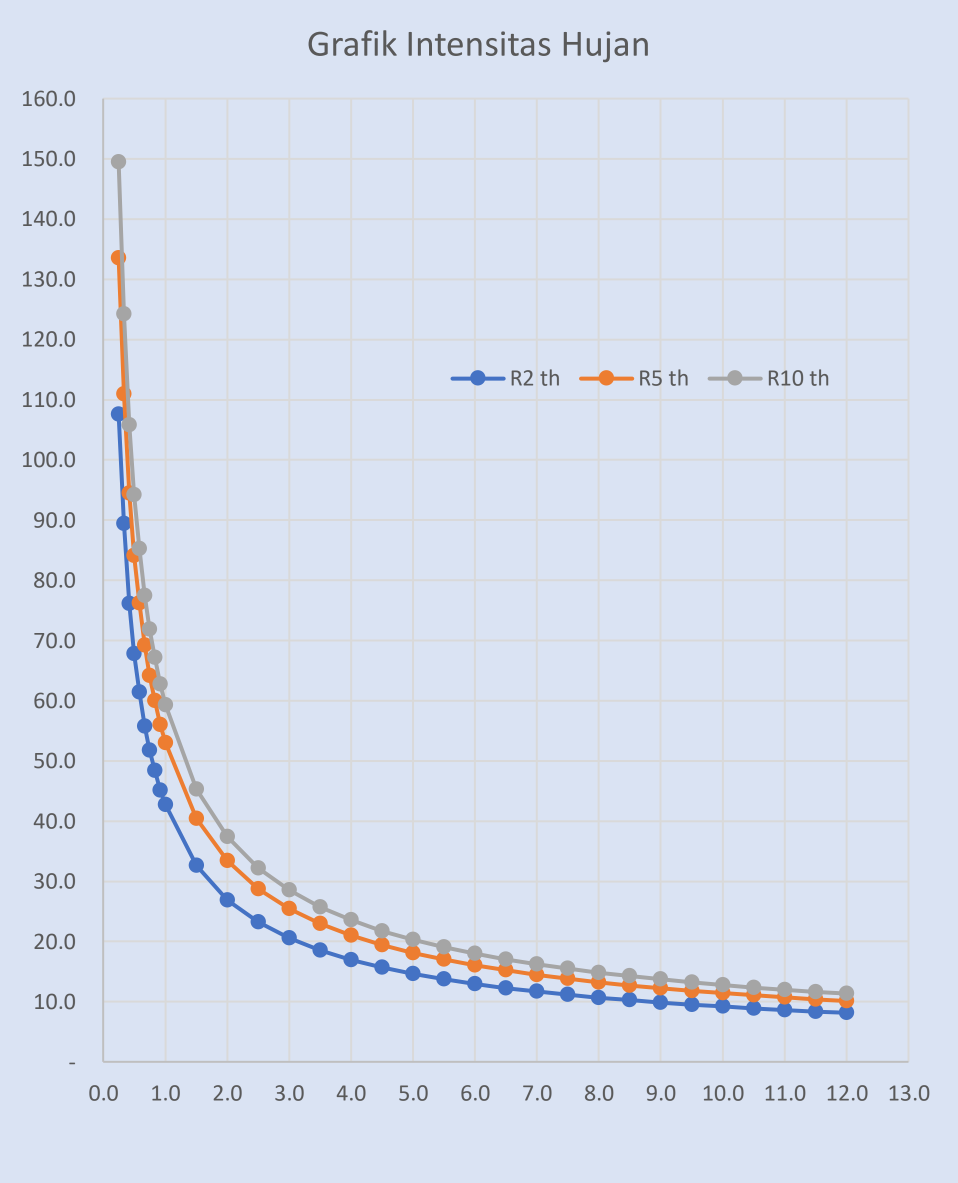
$$\boldsymbol{I = \frac{R24}{24}\left(\frac{24}{t}\right)^{2/3} = \frac{123,160}{24}\left(\frac{24}{1}\right)^{2/3} = 42,7\ mm.jam}$$

Dengan mengubah variabel t untuk masing-masing curah hujan (R24) untuk periode ulang 2 tahun, R24 = 152,805 mm/jam untuk periode ulang 5 tahun dan R24 = 171,080 mm/jam untuk periode ulang 10 tahun, maka hasilnya adalah sebagai berikut berikut :

Dengan interpolasi atau durasi pada Tabel Intesitas Durasi dan Frekuensi (metode mononobe) pada periode ulang 10 tahun dan waktu konsentrasi 0,75 didapatkan Intensitas untuk 10 tahun = 71,8 mm/jam.

Kurva Intensitas Durasi dan rekuensi Hujan

Periode Ulang (thn)	2	5	10
CH Rencana	123,2	152,8	171,1
Durasi hujan (jam)	R2	R5	R10
0,25	107,6	133,5	149,5
0,33	89,4	110,9	124,2
0,42	76,1	94,5	105,8
0,50	67,8	84,1	94,1
0,58	61,4	76,2	85,3
0,67	55,8	69,2	77,5
0,75	**51,7**	**64,2**	**71,8**
0,83	48,3	60,0	67,2
0,92	45,1	56,0	62,7
1,00	42,7	53,0	59,3
1,50	32,6	40,4	45,3
2,00	26,9	33,4	37,4
2,50	23,2	28,8	32,2
3,00	20,5	25,5	28,5
3,50	18,5	23,0	25,7
4,00	16,9	21,0	23,5
4,50	15,7	19,4	21,8
5,00	14,6	18,1	20,3
5,50	13,7	17,0	19,0
6,00	12,9	16,0	18,0
6,50	12,3	15,2	17,0
7,00	11,7	14,5	16,2
7,50	11,1	13,8	15,5
8,00	10,7	13,2	14,8
8,50	10,3	12,7	14,2
9,00	9,9	12,2	13,7
9,50	9,5	11,8	13,2
10,00	9,2	11,4	12,8
10,50	8,9	11,0	12,4
11,00	8,6	10,7	12,0
11,50	8,4	10,4	11,6
12,00	8,1	10,1	11,3



Dengan mengambil nilai koefisien aliran di daerah industri padat C = 0,75 dan luas daerah tangkapan (A) = 44 ha didapat besarnya debit puncak aliran untuk periode ulang 10 tahun (Q10) adalah :

Qp = 0,00278 C.I.A m3/det

= 0,00278 * 0,75*71,8*44

= 6,58 m3/det

TERIMA KASIH

Beberapa asumsi dasar untuk menggunakan metode rasional adalah :

1. Curah hujan terjadi dengan intensitas yang tetap dalam jangka waktu tertentu, setidaknya sama dengan waktu konsentrasi.
2. Limpasan langsung mencapai maksimum ketika durasi hujan dengan intensitas tetap sama dengan waktu konsentrasi.
3. Koefisien *run off* dianggap tetap selama durasi hujan.
4. Luas DAS tidak berubah selama durasi hujan.

(Wanielista, 1990).

kesederhanaanya. Metoda ini dipakai untuk DAS yang kecil. Metoda ini juga menunjukkan parameter-parameter yang dipakai metoda perkiraan banjir lainnya yaitu koefisien *run off*, intensitas hujan, dan luas DAS. Kurva frekuensi intensitas-lamanya dipakai untuk perhitungan limpasan (*run off*) dengan rumus rasional dan untuk perhitungan debit puncak. Luas DAS untuk metoda rasional kurang dari 81 Ha (Dumairy, 1992).

Dalam analisis frekuensi diperlukan seri data hujan yang diperoleh dari pos penakar hujan, baik yang manual maupun yang otomatis. Analisis frekuensi ini didasarkan pada sifat statistik data kejadian yang telah lalu untuk memperoleh probabilitas besaran hujan di masa yang akan datang. Dengan anggapan bahwa sifat statistik kejadian hujan yang akan datang masih sama dengan sifat statistik kejadian hujan masa lalu (Suripin, 2004).

METODE RASIONAL

Pemodelan Debit Puncak Aliran

Muhamad Komarudin

METODE RASIONAL

Implementasi Metode Rasional

- Menggambarkan hubungan antara debit limpasan dengan besar curah hujan
- Menentukan debit puncak banjir / limpasan kala ulang hujan tertentu bagi saluran-saluran dengan daerah aliran kecil, kira-kira 5000 ha.
- Curah hujan diasumsikan merata di daerah tangkapannya

Metode rasional ini dapat dinyatakan secara aljabar dengan persamaan sebagai berikut (Subarkah, 1980) :

$$Q_p = 0,00278\ C.I.A$$

Keterangan:

Qp = debit puncak banjir (m3/s);
C = koefisien limpasan;
I = intensitas hujan selama waktu konsentrasi (mm/jam);
A = luas daerah aliran (Ha).

Menghitung Koefisien Limpasan (C)

Koefisien *runoff* untuk drainase perkotaan sangat dipengaruhi oleh daerah kedap air dan dirumuskan seperti berikut :

$C = 0,9\, I_m + (1 - I_m) C_p$

Keterangan:

C_p = koefisien limpasan *runoff* untuk daerah tidak kedap air;

I_m adalah rasio kedap air.

$$I_m = \frac{A_{KEDAPAIR}}{A_{TOTAL}}$$

Tata guna lahan	Karakteristik	C	Im (%)	Keterangan
Pusat perbelanjaan dan perkantoran		0,90	100	
Industri	Bangunan penuh	0,80	80	Berkurang untuk bangunan tidak penuh
Pemukiman (kepadatan menengah – tinggi)	20 rmh/ha 30 rmh/ha 40 rmh/ha 60 rmh/ha	0,48 0,55 0,65 0,75	30 40 60 75	Bandingkan daerah kedap air dengan daearah lain
Pemukiman (kepadatan rendah)	10 rmh/ha	0,40	< 20	CN =85 (*Curve Number*)
Taman	Daerah datar	0,30	0	
Pedesaan	Tanah berpasir Tanah berat (heavy soil) Daerah irigasi		0 0 0	C = 0,20; CN = 60 C = 0,35; CN = 75 C = 0,50; CN = 85

Sumber: *USDepartment of Agriculture, Natural Resources Conservation Service, June 1986*
CATATAN : CN : *Curve Number* dari SCS

Menghitung Koefisien Limpasan (C)

Untuk C komposit dapat dihitung dengan persamaan seperti berikut :

$$C_k = \frac{C_1.A_1 + C_2.A_2 + \ldots + C_n.A_n}{A_{TOTAL}}$$

Menghitung Koefisien Limpasan (C)

Koefisien Limpasan (C) berdasarkan penggunaan lahan dapat pula ditentukan dengan menggunakan Tabel di bawah ini :

Karakteristik tanah	Tata guna lahan	Koefisien Limpasan (C)
Campuran pasir dan/ atau campuran kerikil	Pertanian	0,20
	Padang rumput	0,15
	Hutan	0,10
Geluh dan sejenisnya	Pertanian	0,40
	Padang rumput	0,35
	Hutan	0,30
Lempung dan sejenisnya	Pertanian	0,50
	Padang rumput	0,45
	Hutan	0,40

Jenis Daerah	Koefisien Aliran	Kondisi Permukaan	Koefisien Aliran
Daerah Perdagangan		Jalan Aspal	
Kota	0,70-0,95	Aspal dan beton	0,75-0,95
Sekitar kita	0,50-0,70	Batu bata dan batako	0,70-0,85
Daerah Pemukiman		Atap Rumah	0,70-0,95
Satu rumah	0,30-0,50	Halaman berumput, tanah pasir	
Banyak Rumah,terpisah	0,40-0,60	Datar, 2%	0,05-0,10
Banyak Rumah, rapat	0,60-0,75	Rata-rata,2-7 %	0,10-0,15
Pemukiman, pinggiran Kota	0,25-0,40	Curam, 7 % atau lebih	0,15-0,20
Apartemen	0,50-0,70		
Daerah Industri		Halaman berumput,tanah pasir padat	
Ringan	0,50-0,80	Datar, 2 %	0,13-0,17
Padat	0,60-0,90	Rata-Rata, 2-7 %	0,18-0,22
Lapangan, kuburan dan sejenisnya	0,10-0,25	Curam, 7 % atau lebih	0,25-0,35
Halaman, jalan kereta api dan sejenisnya	0,20-0,35		
Lahan tidak terpelihara	0,10-0,30		

Menghitung Waktu Konsentrasi (tc)

1) Cara menghitung t_c, Kirpich (1940)

$$t_c = 0{,}0195 \; L^{0,77} \; S^{-0,385}$$

Keterangan:

t_c = waktu dalam menit;

L = panjang lereng dalam; atau Panjang lintasan air dari titik terjauh

sampai titik tinjau (m’)

S = kemiringan lereng rata-rata (m’/m’).

2) Cara menghitung tc,Izzard (1994)

$$t_C = \frac{526,4 * K * L}{i^{2/3}} \quad \text{menit} \rightarrow \text{untuk } i * L < 3871$$

keterangan :

L adalah panjang aliran di lahan (m)

i adalah intensitas hujan (mm/jam)

$$K = \frac{2,756 * 10^{-4} * i + C_r}{S^{2/3}}$$

Keterangan:

L adalah panjang aliran di lahan (*overland flow distance*), dalam m;

i adalah intensitas hujan,dalam mm/jam;

S adalah slope (m/m);

Cr adalah koefisien runoff / aliran (Tabel)

Aspal Halus	**0,0070**
Aspal dan Perkerasan Pasir	**0,0075**
Atap	**0,0082**
Beton	**0,0120**
Aspal dan Perkerasan Krikil	**0,0170**
Rumput	**0,0460**
Alang-Alang	**0,0600**

3) Cara menghitung tc, Kerby (1959)

$$t_C = 1{,}44 * \left(L * n * S^{-0{,}5}\right)^{0{,}467} \text{ menit} \rightarrow \text{untuk } L < 365 \text{ m}$$

Keterangan:

L adalah panjang aliran (m);

S adalah slope (m/m);

n adalah koefisien kekasaran. (Tabel)

Tabel Besarnya koefisien kekasaran (n)

Sumber: Chin, 2000

Paving halus	0,02
Tanah terbuka	0,1
Rumput gersang / tanah terbuka	0,3
Rumput sedang	0,4
Hutan meranggas	0,6
Rumput lebat	0,8

4) Cara menghitung tc, FAA

Daerah perkotaan dengan panjang aliran antara 60 m – 100 m dirumuskan :

$$t_C = \frac{0{,}552 * \left[1{,}8 * (1{,}1 - C) * L^{0{,}5}\right]}{S^{1/3}} \text{ menit.}$$

Keterangan:

C adalah koefisien runoff;

L adalah panjang aliran di lahan (m);

S adalah kemiringan lahan (%).

5) Waktu Konsentrasi di saluran:

$$t_d = \frac{L}{60 * V} \text{ menit}$$

Keterangan:

L adalah panjang saluran (m);

V adalah kecepatan aliran rata-rata (m/s).

Menghitung Intensitas Hujan (I)

Rumus Mononobe :

- Intensitas curah hujan adalah besarnya jumlah hujan yang turun yang dinyatakan dalam tinggi curah hujan atau volume hujan tiap satuan waktu.
- Besarnya intensitas hujan berbeda-beda, tergantung dari lamanya curah hujan dan frekuensi kejadiannya

$$I = \frac{R24}{24}\left(\frac{24}{t}\right)^{2/3}$$

Dimana :

I : Intensitas hujan (mm/jam)

R24 : curah hujan harian maksimum untuk kala ulang T (mm/jam)

t : Lama hujan (jam)

Contoh perhitungan debit puncak aliran (Qp)

Suatu wilayah Kawasan industri memiliki karakteristik daerah tangkapan sebagai berikut :

- Luas Tangkapan (A) : 44 HA
- Panjang Aliran (L) : 1050 m
- Rata-rata Kemiringan (S) : 0,0021 m / m
- Penguunaan Lahan : Kawasan Industri padat

  Dari table diperoleh nilai Koefisien Limpasan (C) : 0,75
- Curah Hujan periode ulang 10 tahun : 171,1 mm
- Waktu konsentrasi menurut metode Kirpich :

  $tc = 0{,}0195\ L^{0{,}77}\ S^{-0{,}385}$ (menit)

  = 0,0195*1050^0,77*0,0021^-0,386

  = 44,67 menit = 0,75 jam

Menghitung Intensitas Hujan (I)

Diketahui curah hujan rencana (R) sebesar 123.160 mm pada kala ulang 2 tahun, dengan lama hujan (t) adalah 1 jam. maka perhitungan Intensitas adalah sebagai berikut:

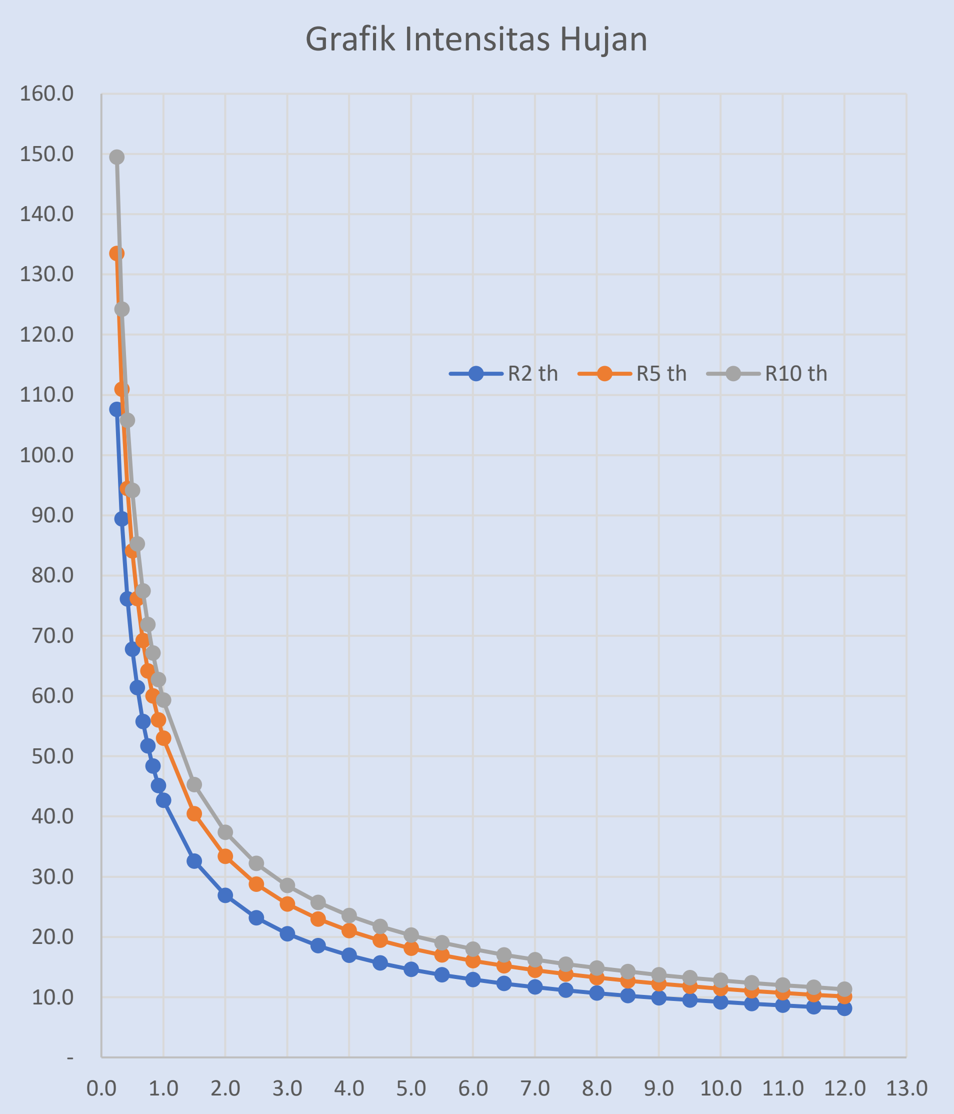
$$\boldsymbol{I = \frac{R24}{24}\left(\frac{24}{t}\right)^{2/3} = \frac{123,160}{24}\left(\frac{24}{1}\right)^{2/3} = 42,7\ mm.jam}$$

Dengan mengubah variabel t untuk masing-masing curah hujan (R24) untuk periode ulang 2 tahun, R24 = 152,805 mm/jam untuk periode ulang 5 tahun dan R24 = 171,080 mm/jam untuk periode ulang 10 tahun, maka hasilnya adalah sebagai berikut berikut :

Dengan interpolasi atau durasi pada Tabel Intesitas Durasi dan Frekuensi (metode mononobe) pada periode ulang 10 tahun dan waktu konsentrasi 0,75 didapatkan Intensitas untuk 10 tahun = 71,8 mm/jam.

Kurva Intensitas Durasi dan rekuensi Hujan

Periode Ulang (thn)	2	5	10
CH Rencana	123,2	152,8	171,1
Durasi hujan (jam)	R2	R5	R10
0,25	107,6	133,5	149,5
0,33	89,4	110,9	124,2
0,42	76,1	94,5	105,8
0,50	67,8	84,1	94,1
0,58	61,4	76,2	85,3
0,67	55,8	69,2	77,5
0,75	**51,7**	**64,2**	**71,8**
0,83	48,3	60,0	67,2
0,92	45,1	56,0	62,7
1,00	42,7	53,0	59,3
1,50	32,6	40,4	45,3
2,00	26,9	33,4	37,4
2,50	23,2	28,8	32,2
3,00	20,5	25,5	28,5
3,50	18,5	23,0	25,7
4,00	16,9	21,0	23,5
4,50	15,7	19,4	21,8
5,00	14,6	18,1	20,3
5,50	13,7	17,0	19,0
6,00	12,9	16,0	18,0
6,50	12,3	15,2	17,0
7,00	11,7	14,5	16,2
7,50	11,1	13,8	15,5
8,00	10,7	13,2	14,8
8,50	10,3	12,7	14,2
9,00	9,9	12,2	13,7
9,50	9,5	11,8	13,2
10,00	9,2	11,4	12,8
10,50	8,9	11,0	12,4
11,00	8,6	10,7	12,0
11,50	8,4	10,4	11,6
12,00	8,1	10,1	11,3



Dengan mengambil nilai koefisien aliran di daerah industri padat C = 0,75 dan luas daerah tangkapan (A) = 44 ha didapat besarnya debit puncak aliran untuk periode ulang 10 tahun (Q10) adalah :

$$Qp = 0{,}00278\ C.I.A\ \ m3/det$$

$$= 0{,}00278 * 0{,}75*71{,}8*44$$

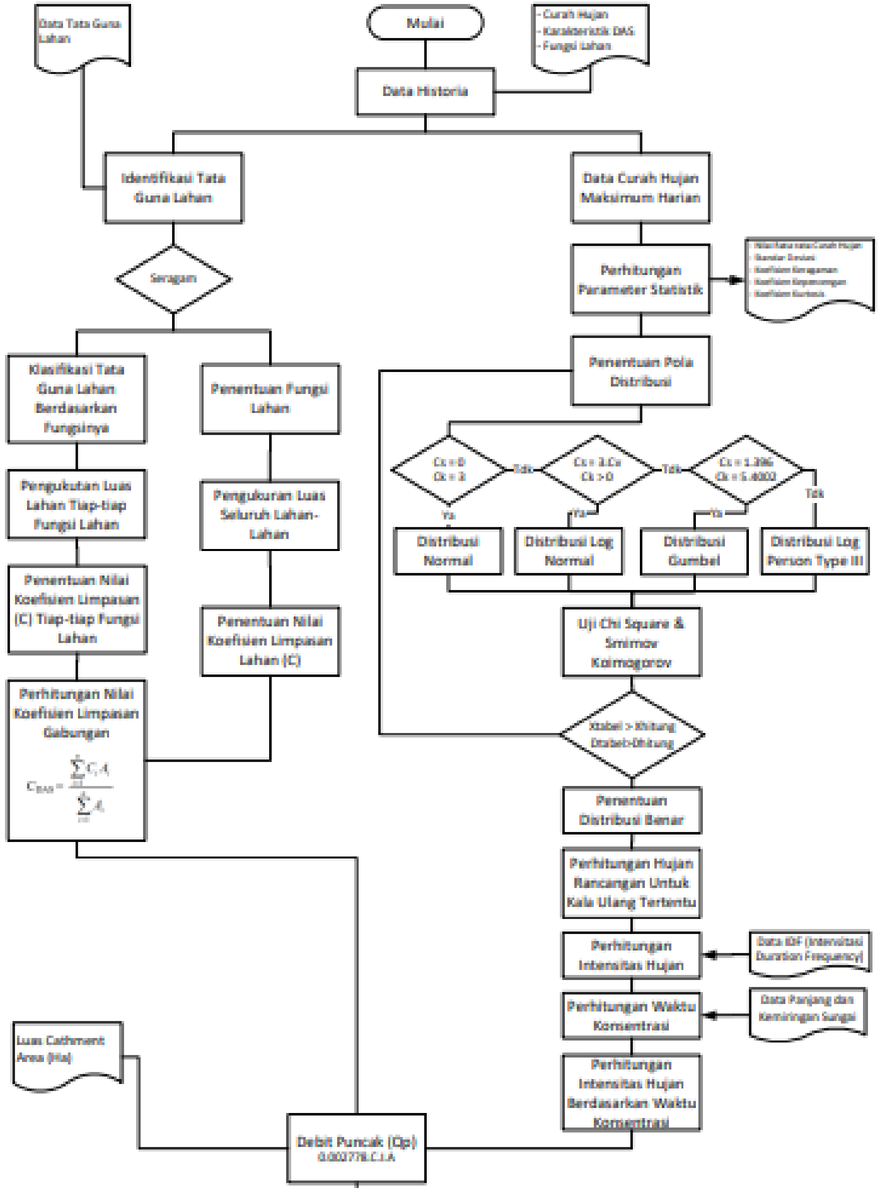
$$\mathbf{= 6{,}58\ m3/det}$$

TERIMA KASIH

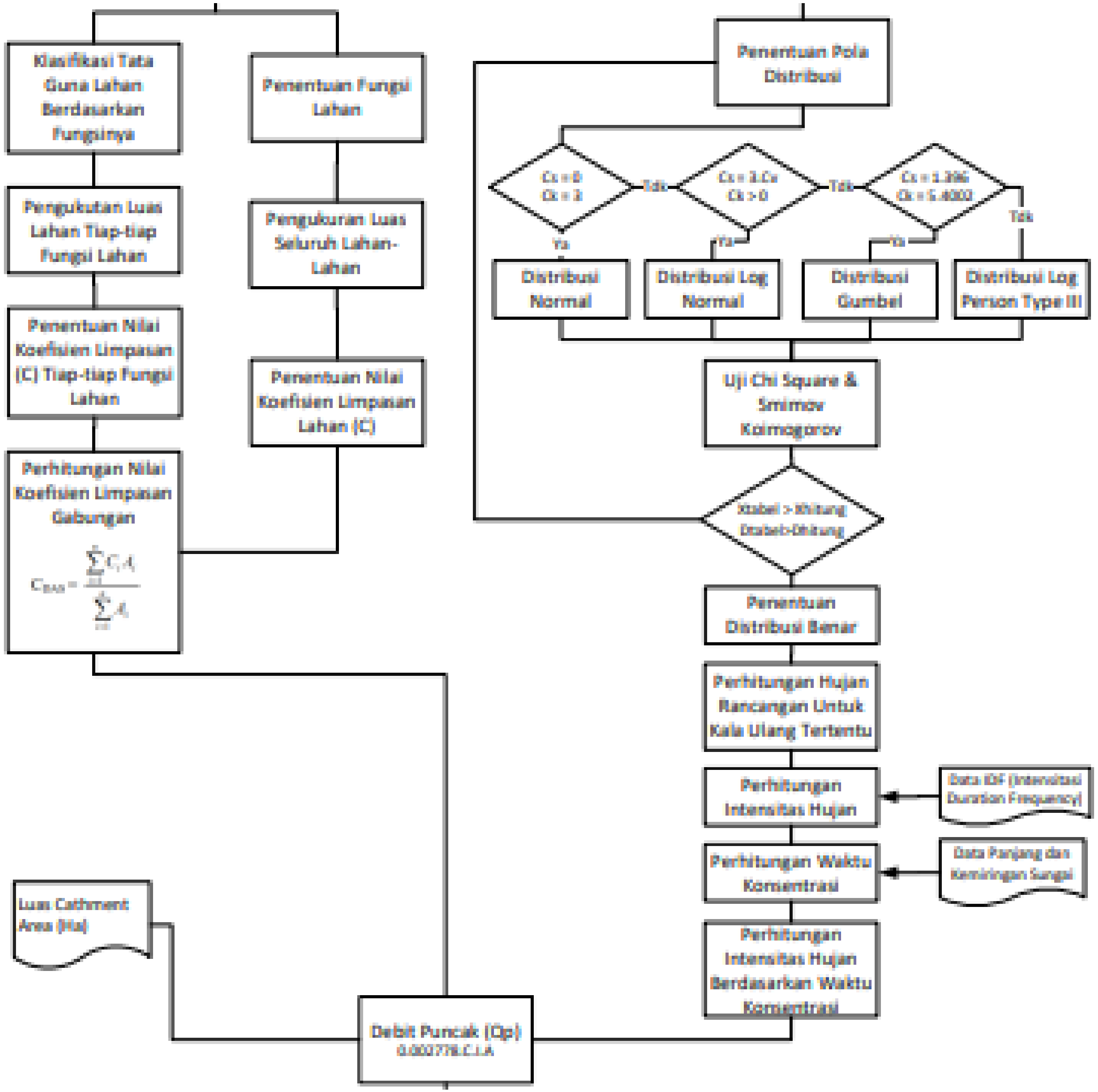
IMPLEMENTASI METODE RASIONAL

Pemodelan Debit Puncak Aliran

Muhamad Komarudin



Mulai
Data Tata Guna Lahan
- Curah Hujan
- Karakteristik DAS
- Fungsi Lahan
Data Historis
Identifikasi Tata Guna Lahan
Data Curah Hujan Maksimum Harian
Seragam
Perhitungan Parameter Statistik
Penentuan Pola Distribusi
Klasifikasi Tata Guna Lahan Berdasarkan Fungsinya
Penentuan Fungsi Lahan
Cs ≈ 0 Ck ≈ 3
Tdk
Cs ≈ 3.Cv Ck > 0
Cs ≈ 1.396 Ck ≈ 5.4002
Ya
Pengukutan Luas Lahan Tiap-tiap Fungsi Lahan
Pengukuran Luas Seluruh Lahan-Lahan
Distribusi Normal
Distribusi Log Normal
Distribusi Gumbel
Distribusi Log Person Type III
Penentuan Nilai Koefisien Limpasan (C) Tiap-tiap Fungsi Lahan
Penentuan Nilai Koefisien Limpasan Lahan (C)
Uji Chi Square & Smirnov Kolmogorov
Perhitungan Nilai Koefisien Limpasan Gabungan
Xtabel > Xhitung Dtabel>Dhitung
Penentuan Distribusi Benar
Perhitungan Hujan Rancangan Untuk Kala Ulang Tertentu
Perhitungan Intensitas Hujan
Data IDF (Intensitasi Duration Frequency)
Perhitungan Waktu Konsentrasi
Data Panjang dan Kemiringan Sungai
Luas Cathment Area (Ha)
Perhitungan Intensitas Hujan Berdasarkan Waktu Konsentrasi
Debit Puncak (Qp) 0.002778.C.I.A



Klasifikasi Tata Guna Lahan Berdasarkan Fungsinya
Penentuan Fungsi Lahan
Penentuan Pola Distribusi
Pengukutan Luas Lahan Tiap-tiap Fungsi Lahan
Pengukuran Luas Seluruh Lahan-Lahan
Cs ≈ 0 Ck ≈ 3
Tdk
Cs ≈ 3.Cv Ck > 0
Cs ≈ 1.396 Ck ≈ 5.4002
Ya
Distribusi Normal
Distribusi Log Normal
Distribusi Gumbel
Distribusi Log Person Type III
Penentuan Nilai Koefisien Limpasan (C) Tiap-tiap Fungsi Lahan
Penentuan Nilai Koefisien Limpasan Lahan (C)
Uji Chi Square & Smirnov Kolmogorov
Perhitungan Nilai Koefisien Limpasan Gabungan
Xtabel > Xhitung Dtabel>Dhitung
Penentuan Distribusi Benar
Perhitungan Hujan Rancangan Untuk Kala Ulang Tertentu
Perhitungan Intensitas Hujan
Data IDF (Intensitasi Duration Frequency)
Perhitungan Waktu Konsentrasi
Data Panjang dan Kemiringan Sungai
Luas Cathment Area (Ha)
Perhitungan Intensitas Hujan Berdasarkan Waktu Konsentrasi
Debit Puncak (Qp) 0.002778.C.I.A

Beberapa asumsi dasar untuk menggunakan metode rasional adalah :

1. Curah hujan terjadi dengan intensitas yang tetap dalam jangka waktu tertentu, setidaknya sama dengan waktu konsentrasi.
2. Limpasan langsung mencapai maksimum ketika durasi hujan dengan intensitas tetap sama dengan waktu konsentrasi.
3. Koefisien *run off* dianggap tetap selama durasi hujan.
4. Luas DAS tidak berubah selama durasi hujan.

(Wanielista, 1990).

Data kondisi DAS Belawan yang diperoleh dari Dinas Pengairan Kota Medan dan Yayasan Leuser Indonesia adalah sebagai :

Luas total daerah pengaliran sungai Belawan (A) = 439,37 km^2

Panjang sungai Belawan = 65 km

Kemiringan = 0,00798 m/m

Tabel 3. Data penggunaan lahan pada DAS Belawan

Jenis penutup tanah	A (km^2)
Hutan Primer	2.37
Hutan Mangrove	6.92
Hutan Sekunder	5.26
Tanah Terbuka/padang rumput	0.70
Kebun Campuran	387.73
Kelapa Sawit	8.52
Sawah Irigasi	3.23
Daerah Perkotaan	15.46
Total	430.19

Sumber : Yayasan Leuser Indonesia

Tabel 4. Data curah hujan harian maksimum tahunan periode 1985-2006

No	Rmax
1	45
2	60
3	64
4	67
5	68
6	68
7	71
8	71
9	71
10	72
11	75
12	76
13	80
14	82
15	94
16	94
17	95
18	96
19	100
20	103
21	118
22	155

PENENTUAN POLA DISTRIBUSI HUJAN

Data Curah Hujan Harian Maksimum tahunan periode 1985-2006

No	R_{Max}
1	45
2	60
3	64
4	67
5	68
6	68
7	71
8	71
9	71
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11	75
12	76
13	80
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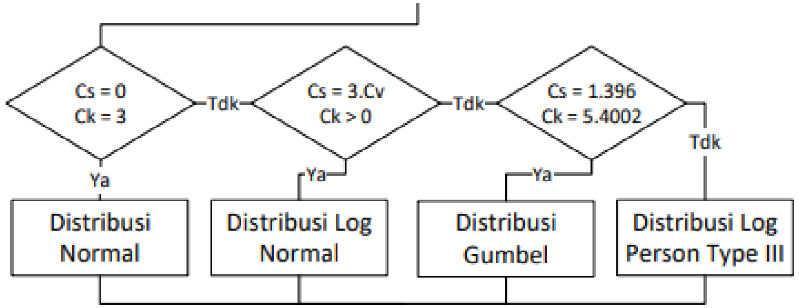
15	94
16	94
17	95
18	96
19	100
20	103
21	118

Hasil Perhitungan Parameter Statistik

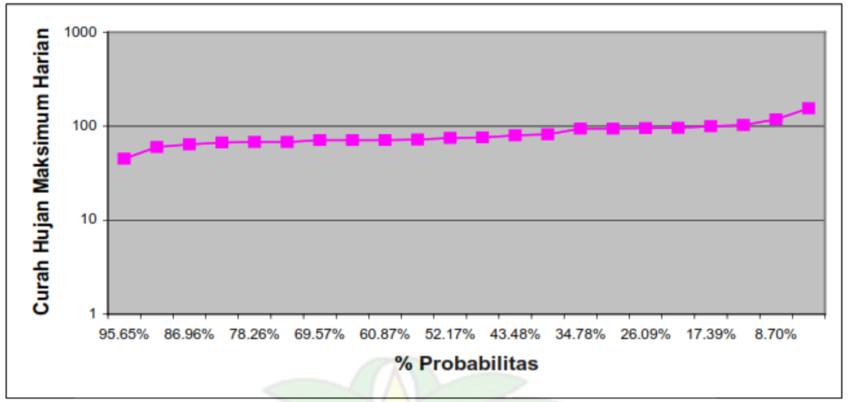
Rata-rata	$\bar{x} = 82{,}95$
Simpangan Baku	$s = 23{,}27$
Koefisien Variasi	Cv = 0,28
Koefisien Skewness	Cs = 1,43
Koefisien Kurtosis	Ck = 3,33

Hitungan parameter statistic yang diperoleh tersebut ditetapkan bahwa jenis distribusi yang cocok dengan sebaran data curah hujan harian maksimum di wilayah studi adalah **Distribusi Log Pearson Type III** untuk menghitung curah hujan rancangan dengan berbagai kala ulang. Hal ini ditunjukan oleh nilai parameter statistic yang diperoleh tidak mengikuti pola distribusi untuk ketiga metode lainnya dan penggambaran garis teoritiknya berupa garis lengkung.



Menurut Jayadi, 2000 , Ciri Distribusi Log Person Tipe III yaitu :

1. Jika tidak menunjukkan sifat-sifat seperti ketida distribusi yaitu distribusi Gumbel, Normal maupun Log Normal
2. Garis Teoritisnya Probabilitasnya berupa garis lengkung



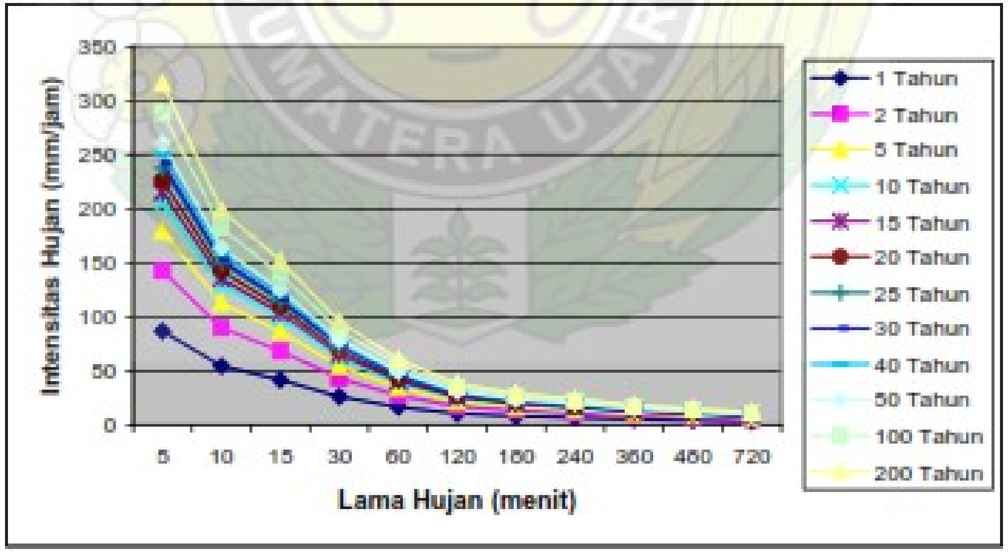
Uji Kecocokan	Nilai Tabel		Nilai Hitung
Chi-Square	3,841	>	2,0
Smirnov-Kolmogorov	0,276	>	0,1048

Nilai Intensitas Hujan

Kala Ulang (Tahun)	Hujan Rancangan (mm)
1	47.38
2	78.61
5	98.97
10	112.93
15	118.66
20	124.82
25	131.13
30	133.54
40	138.52
50	143.65
100	159.55
200	174.5

Tabel 9. Intensitas hujan jam-jaman

T	Kala Ulang											
	1	2	5	10	15	20	25	30	40	50	100	200
(menit)												
5	86.52	142.34	179.21	204.49	214.86	226.02	237.44	241.81	250.83	260.12	288.91	315.98
10	54.51	89.69	112.92	128.84	135.38	142.41	149.61	152.36	158.04	163.89	182.03	199.09
15	41.61	68.46	86.20	98.35	103.34	108.71	114.20	116.30	120.64	125.11	138.96	151.98
30	26.23	43.15	54.32	61.99	65.13	68.51	71.98	73.30	76.03	78.85	87.58	95.78
60	16.53	27.19	34.24	39.07	41.05	43.18	45.36	46.20	47.92	49.70	55.20	60.37
120	10.42	17.14	21.58	24.62	25.87	27.21	28.59	29.12	30.20	31.32	34.79	38.05
180	7.95	13.08	16.47	18.80	19.75	20.77	21.82	22.23	23.05	23.91	26.55	29.04
240	6.57	10.80	13.60	15.52	16.31	17.15	18.02	18.35	19.03	19.74	21.92	23.98
360	5.01	8.25	10.38	11.85	12.45	13.09	13.76	14.01	14.53	15.07	16.74	18.30
480	4.14	6.81	8.57	9.78	10.28	10.81	11.36	11.57	12.00	12.44	13.82	15.11
720	3.16	5.20	6.54	7.47	7.84	8.25	8.67	8.83	9.16	9.50	10.55	11.54



Gambar 4. Kurva *IDF (Intensity Duration Frequency)*

Koefisien Run Off

Jenis penutup tanah	A (km^2)	C	C x A
Hutan Primer	2.37	0.02	0.0474
Hutan Mangrove	6.92	0.02	0.1384
Hutan Sekunder/sangat terdegradasi	5.26	0.05	0.263
Tanah Terbuka/padang rumput	0.7	0.2	0.14
Kebun Campuran	387.73	0.2	38.773
Kelapa Sawit	8.52	0.4	3.408
Sawah Irigasi	3.23	0.15	0.4845
Daerah Perkotaan	15.46	0.9	9.276
Total	430.19	1.64	95.9413
Nilai C	**0,2230**		

Berdasarkan perhitungan di atas dapat dinyatakan bahwa pada kala ulang 1 tahun selama durasi hujan (waktu konsentrasi) 10,60 jam dengan intensitas hujan 3,5 mm/jam seluas 439,37 km^2 maka debit puncak yang diperoleh pada DAS Belawan sebesar 95,27 m^3/detik. Debit puncak yang diperoleh dapat dijadikan sebagai bahan dasar untuk perencanaan bangunan pengendali banjir, dimana dibangun suatu bangunan pengendali banjir yang dapat menampung debit puncak suatu aliran air sehingga dapat menghemat biaya dan waktu dalam pelaksanaan proyek pembangunan.

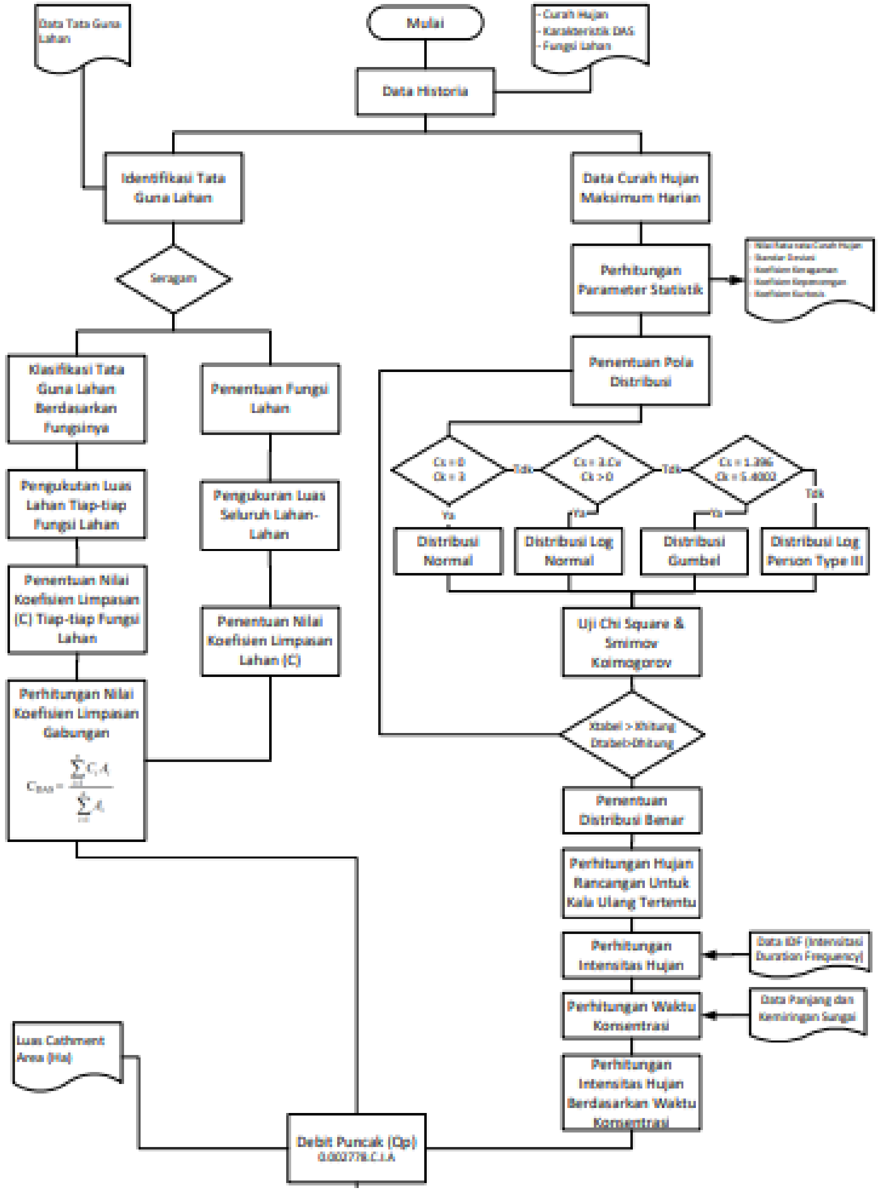
Tabel 11. Debit puncak di DAS Belawan

Kala Ulang (tahun)	Intensitas (mm/jam)	Debit Puncak (m^3/detik)
1	3.5	95.27
2	5.76	156.78
5	7.25	197.34
10	8.28	225.37
15	8.69	236.53
20	9.15	249.05
25	9.61	261.57
30	9.79	266.47
40	10.15	276.27
50	10.53	286.61
100	11.69	318.19
200	12.79	348.13

IMPLEMENTASI METODE RASIONAL

Penentuan Pola Distribusi Hujan

Muhamad Komarudin



Data Tata Guna Lahan
Mulai
- Curah Hujan
- Karakteristik DAS
- Fungsi Lahan
Data Historis
Identifikasi Tata Guna Lahan
Data Curah Hujan Maksimum Harian
Perhitungan Parameter Statistik
Seragam
Klasifikasi Tata Guna Lahan Berdasarkan Fungsinya
Penentuan Fungsi Lahan
Pengukutan Luas Lahan Tiap-tiap Fungsi Lahan
Pengukuran Luas Seluruh Lahan-Lahan
Penentuan Nilai Koefisien Limpasan (C) Tiap-tiap Fungsi Lahan
Penentuan Nilai Koefisien Limpasan Lahan (C)
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$C_{DAS} = \frac{\sum_{i=1}^{n} C_i A_i}{\sum_{i=1}^{n} A_i}$
Penentuan Pola Distribusi
Cs ≈ 0 Ck ≈ 3
Tdk
Cs ≈ 3.Cv Ck > 0
Tdk
Cs ≈ 1.396 Ck ≈ 5.4002
Tdk
Ya
Ya
Ya
Distribusi Normal
Distribusi Log Normal
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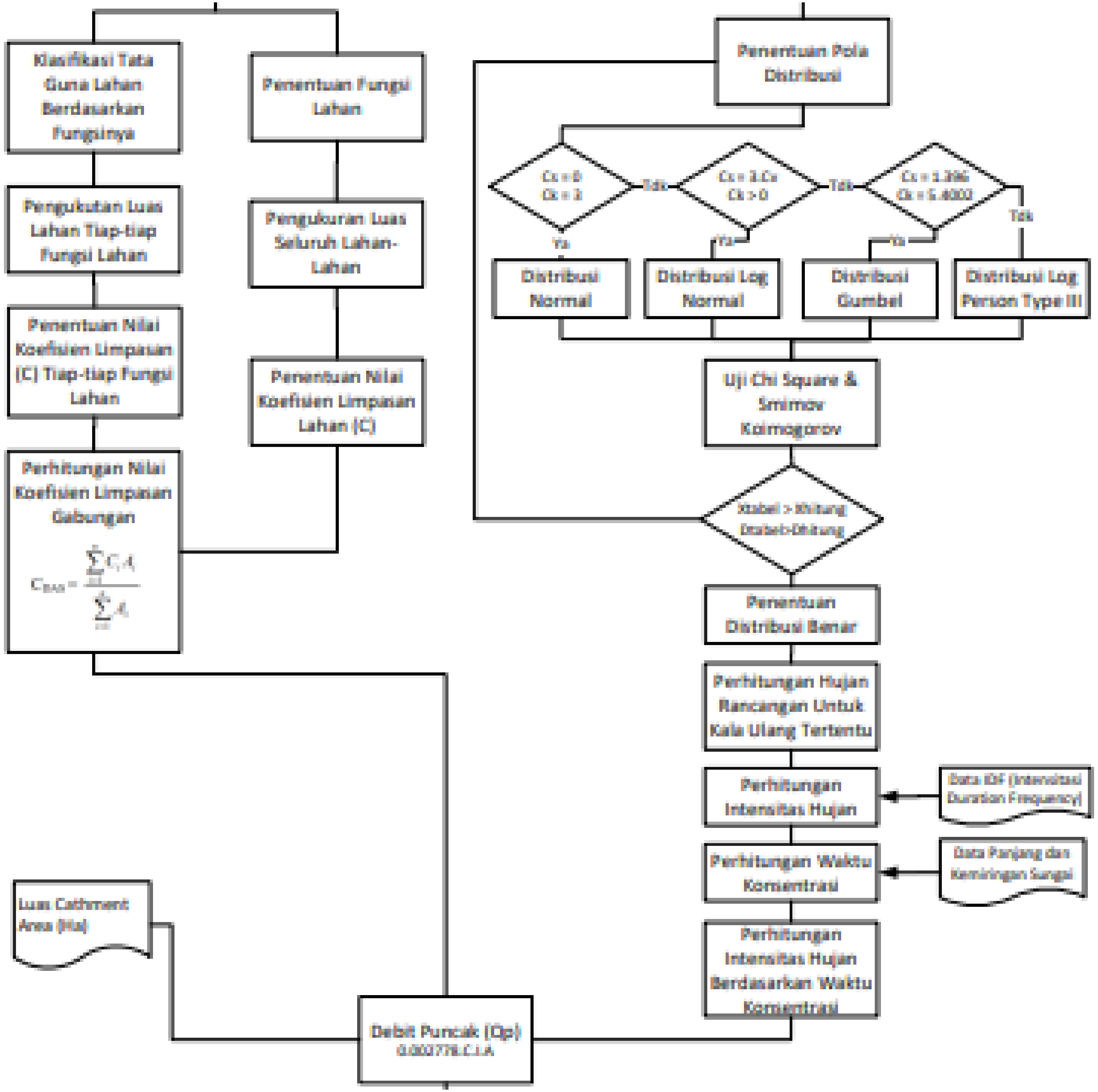

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Distribusi Normal
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(Wanielista, 1990).

ANALISIS DISTRIBUSI FREKWENSI

1. Distribusi Gumbel
2. Distribusi Log Pearson Tipe III
3. Distribusi Normal
4. Distribusi Log Normal

PRINSIP ANALISIS FREKUENSI

Parameter Statistik

- -Hujan rata-rata (x)
- -Standart Deviasi (Sd)
- -koef Skewness(Cs)
- -Koef variasi (Cv)
- -Koef Kurtosis(Ck)

No	Jenis Sebaran	Hasil Perhitungan	Syarat	Keterangan
1	Gumbel	Cs = 0,77	Cs $\leq$ 1,1396	Tidak memenuhi
		Ck = 6,62	Ck $\leq$ 5,4002	
2	Normal	Cs = 0,77	Cs $\approx$ 0	Tidak memenuhi
		Ck = 6,62	Ck $\approx$ 3	
3	Log Normal	Cs = -0,9	Cs $\approx$ 3Cv + Cv^2 = 3	Tidak Memenuhi
		Ck = 6,071	Ck = 5,383	
4	Log Pearson Tipe III	Cs = -0,9	Cs $\neq$ 0	Memenuhi

Pemilihan sebaran

Pengujian sebaran

- Cara Grafis
- *Smirnov-Kolmogorv*
- Uji Chi Kuadrat

PARAMETER STATISTIK UNTUK ANALISIS FREKUENSI

Tabel 1. Parameter statistik analisis frekuensi

Parameter	Sampel
Rata-rata	$X = \frac{1}{n}\sum_{i=1}^{n} X_i$
Simpangan Baku	$S = \left[\frac{1}{n-1}\sum_{i=1}^{n}\left(\log X_i - \log \overline{X}\right)^2\right]^{1/2}$
Koefisien Variasi	$Cv = \frac{S}{\overline{v}}$
Koefisien Skewness	$Cs = \frac{n\sum_{i=1}^{n}\left(X_i - \overline{X}\right)^3}{(n-1)(n-2)s^3}$
Koefisien Kurtosis	$Ck = \frac{n^2\sum_{i=1}^{n}\left(X_i - \overline{X}\right)^4}{(n-1)(n-2)(n-3)s^4}$

Sumber: Singh, 1992.

Tabel 4. Data curah hujan harian maksimum tahunan periode 1985-2006

No	Rmax
1	45
2	60
3	64
4	67
5	68
6	68
7	71
8	71
9	71
10	72
11	75
12	76
13	80
14	82
15	94
16	94
17	95
18	96
19	100
20	103
21	118
22	155

Untuk menghitung curah hujan rencana dengan metode sebaran Gumbel digunakan persamaan distribusi frekuensi empiris sebagai berikut *(CD.Soemarto, 1999)* :

$$X_T = \overline{X} + \frac{Sd}{Sn}(Y_T - Yn)$$

$$Sd = \sqrt{\frac{\sum (X_i - \overline{X})^2}{n-1}}$$

dimana :

X_T = nilai hujan rencana dengan data ukur T tahun.

= nilai rata - rata hujan

Sd = standar deviasi (simpangan baku)

Y_T = nilai reduksi variat *(reduced variate)* dari variabel yang diharapkan terjadi pada periode ulang T tahun, dapat dilihat pada Tabel 1

Y_n = nilai rata-rata dari reduksi variat *(reduce mean)* nilainya tergantung dari jumlah data (n), dapat dilihat pada Tabel 2

S_n = deviasi standar dari reduksi variat *(reduced standart deviation)* nilainya tergantung dari jumlah data (n), dapat dilihat pada Tabel 3

Tabel 1. nilai reduksi variat (YT)

Periode Ulang (Tahun)	Reduced Variate
2	0,3665
5	1,4999
10	2,2502
20	2,9606
25	3,1985
50	3,9019
100	4,6001
200	5,2960
500	6,2140
1000	6,9190
5000	8,5390
10000	9,9210

Tabel 2. nilai rata-rata dari reduksi variat (Yn)

n	0	1	2	3	4	5	6	7	8	9
10	0,4952	0,4996	0,5035	0,5070	0,5100	0,5128	0,5157	0,5181	0,5202	0,5220
20	0,5236	0,5252	0,5268	0,5283	0,5296	0,5300	0,5820	0,5882	0,5343	0,5353
30	0,5363	0,5371	0,5380	0,5388	0,5396	0,5400	0,5410	0,5418	0,5424	0,5430
40	0,5463	0,5442	0,5448	0,5453	0,5458	0,5468	0,5468	0,5473	0,5477	0,5481
50	0,5485	0,5489	0,5493	0,5497	0,5501	0,5504	0,5508	0,5511	0,5515	0,5518
60	0,5521	0,5524	0,5527	0,5530	0,5533	0,5535	0,5538	0,5540	0,5543	0,5545
70	0,5548	0,5550	0,5552	0,5555	0,5557	0,5559	0,5561	0,5563	0,5565	0,5567
80	0.5569	0,5570	0,5572	0,5574	0,5576	0,5578	0,5580	0,5581	0,5583	0,5585
90	0,5586	0,5587	0,5589	0,5591	0,5592	0,5593	0,5595	0,5596	0,5598	0,5599
100	0,56									

Tabel 3. deviasi standar dari reduksi variat (Sn)

n	0	1	2	3	4	5	6	7	8	9
10	0,9496	0,9676	0,9833	0,9971	1,0095	1,0206	1,0316	1,0411	1,0493	1,0565
20	1,0628	1,0696	1,0754	1,0811	1,0864	1,0315	1,0961	1,1004	1,1047	1,1080
30	1,1124	1,1159	1,1193	1,1226	1,1255	1,1285	1,1313	1,1339	1,1363	1,1388
40	1,1413	1,1436	1,1458	1,1480	1,1499	1,1519	1,1538	1,1557	1,1574	1,1590
50	1,1607	1,1923	1,1638	1,1658	1,1667	1,1681	1,1696	1,1708	1,1721	1,1734
60	1,1747	1,1759	1,1770	1,1782	1,1793	1,1803	1,1814	1,1824	1,1834	1,1844
70	1,1854	1,1863	1,1873	1,1881	1,1890	1,1898	1,1906	1,1915	1,1923	1,1930
80	1,1938	1,1945	1,1953	1,1959	1,1967	1,1973	1,1980	1,1987	1,1994	1,2001
90	1,2007	1,2013	1,2026	1,2032	1,2038	1,2044	1,2046	1,2049	1,2055	1,2060
100	1,2065									

PARAMETER STATISTIK UNTUK ANALISIS FREKUENSI

Tabel 1. Parameter statistik analisis frekuensi

Parameter	Sampel
Rata-rata	$X = \frac{1}{n}\sum_{i=1}^{n} X_i$
Simpangan Baku	$S = \left[\frac{1}{n-1}\sum_{i-1}^{n}\left(\log X_i - \log \overline{X}\right)^2\right]^{1/2}$
Koefisien Variasi	$Cv = \frac{S}{\overline{v}}$
Koefisien Skewness	$Cs = \frac{n\sum_{i=1}^{n}\left(X_i - \overline{X}\right)^3}{(n-1)(n-2)s^3}$
Koefisien Kurtosis	$Ck = \frac{n^2\sum_{i=1}^{n}\left(X_i - \overline{X}\right)^4}{(n-1)(n-2)(n-3)s^4}$

Sumber: Singh, 1992.

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LOG PERSON TYPE III

$$Y \quad = \bar{Y} + K.S$$

$$\text{Log}\ (X_T) = \overline{\log(X)} + K\ .Sd$$

$$Sd = \sqrt{\frac{\sum_{i=1}^{n}\left\{\log(Xi) - \overline{\log(X)}\right\}^2}{n-1}}$$

Y = Log (X_T) = Nilai Curah Hujan periode ulang T tahun
X = Data curah hujan
$\bar{Y}$ = $\overline{\log(X)}$ = Nilai rata curah hujan logaritmik
Sd = Standar Deviasi
K = Karakteristik distribusi Log Pearson Tipe III (Tabel 4)
Cs = Koefisien Skewnes/ Koefisien Kemencengan
n = Jumlah data hujan

$$\overline{\log(X)} = \frac{\sum_{i=1}^{n}\log(Xi)}{n}$$

$$Cs = \frac{\sum_{i=1}^{n}\left\{\log(Xi) - \overline{\log(X)}\right\}^3}{(n-1)(n-2)Sd^3}$$

HARGA K

Koefisien Kemencengan (Cs)	Periode Ulang Tahun							
	2	5	10	25	50	100	200	1000
	Peluang (%)							
	50	20	10	4	2	1	0,5	0,1
3,0	-0,396	0,420	1,180	2,278	3,152	4,051	4,970	7,250
2,5	-0,360	0,518	1,250	2,262	3,048	3,845	4,652	6,600
2,2	-0,330	0,574	1,284	2,240	2,970	3,705	4,444	6,200
2,0	-0,307	0,609	1,302	2,219	2,912	3,605	4,298	5,910
1,8	-0,282	0,643	1,318	2,193	2,848	3,499	4,147	5,660
1,6	-0,254	0,675	1,329	2,163	2,780	3,388	3,990	5,390
1,4	-0,225	0,705	1,337	2,128	2,706	3,271	3,828	5,110
1,2	-0,195	0,732	1,340	2,087	2,626	3,149	3,661	4,820
1,0	-0,164	0,758	1,340	2,043	2,542	3,022	3,489	4,540
0,9	-0,148	0,769	1,339	2,018	2,498	2,957	3,401	4,395
0,8	-0,132	0,780	1,336	1,993	2,453	2,891	3,312	4,250
0,7	-0,116	0,790	1,333	1,967	2,407	2,824	3,223	4,105
0,6	-0,099	0,800	1,328	1,939	2,359	2,755	3,132	3,960
0,5	-0,083	0,808	1,323	1,910	2,311	2,686	3,041	3,815
0,4	-0,066	0,816	1,317	1,880	2,261	2,615	2,949	3,670
0,3	-0,050	0,824	1,309	1,849	2,211	2,544	2,856	3,525
0.2	-0,033	0,830	1,301	1,818	2,159	2,472	2,763	3,380
0,1	-0,017	0,836	1,292	1,785	2,107	2,400	2,670	3,235
0,0	0,000	0,842	1,282	1,751	2,054	2,326	2,576	3,090
-0,1	0,017	0,836	1,270	1,761	2,000	2,252	2,482	3,950
-0,2	0,033	0,850	1,258	1,680	1,945	2,178	2,388	2,810
-0,3	0,050	0,853	1,245	1,643	1,890	2,104	2,294	2,675
-0,4	0,066	0,855	1,231	1,606	1,834	2,029	2,201	2,540
-0,5	0,083	0,856	1,216	1,567	1,777	1,955	2,108	2,400
-0,6	0,099	0,857	1,200	1,528	1,720	1, 880	2,016	2,275
-0,7	0,116	0,857	1,183	1,488	1,663	1,806	1,926	2,150
-0,8	0,132	0,856	1,166	1,488	1,606	1,733	1,837	2,035
-0,9	0,148	0,854	1,147	1,407	1,549	1,660	1,749	1,910
-1,0	0,164	0,852	1,128	1,366	1,492	1,588	1,664	1,800
-1,2	0,195	0,844	1,086	1,282	1,379	1,449	1,501	1,625
-1,4	0,225	0,832	1,041	1,198	1,270	1,318	1,351	1,465

$$Xt = \bar{X} + Kt.Sd$$

X_T = besarnya curah hujan dengan periode ulang T tahun.

$\overline{X}$ = curah hujan rata-rata (mm)

Sd = Standar Deviasi data hujan harian maksimum

Kt = *Standard Variable* untuk periode ulang t tahun yang besarnya diberikan pada Tabel 6

Tabel 6. Metode Distribusi Normal – Nilai Variabel Reduksi Gauss

No.	Periode Ulang (T) Tahun	Peluang	K_T
1.	1,001	0,999	-3,05
2.	1,250	0,800	-0,84
3.	1,670	0,600	-0,25
4.	2,500	0,400	0,25
5.	2,000	0,500	0
6.	5,000	0,200	0,84
7.	10,000	0,100	1,28
8.	20,000	0,050	1,64
9.	50,000	0,020	2,05
10.	100,000	0,010	2,33

PENENTUAN POLA DISTRIBUSI HUJAN

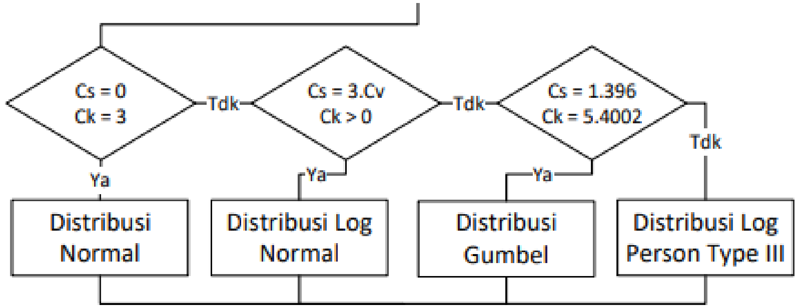
Data Curah Hujan Harian Maksimum tahunan periode 1985-2006

No	R_{Max}
1	45
2	60
3	64
4	67
5	68
6	68
7	71
8	71
9	71
10	72
11	75
12	76
13	80
14	82
15	94
16	94
17	95
18	96
19	100
20	103
21	118

Hasil Perhitungan Parameter Statistik

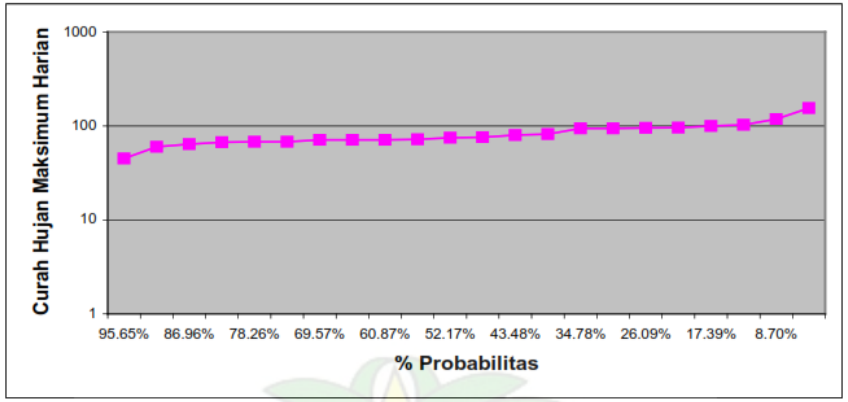
Rata-rata	$\bar{x} = 82{,}95$
Simpangan Baku	$s = 23{,}27$
Koefisien Variasi	Cv = 0,28
Koefisien Skewness	Cs = 1,43
Koefisien Kurtosis	Ck = 3,33

Hitungan parameter statistic yang diperoleh tersebut ditetapkan bahwa jenis distribusi yang cocok dengan sebaran data curah hujan harian maksimum di wilayah studi adalah **Distribusi Log Pearson Type III** untuk menghitung curah hujan rancangan dengan berbagai kala ulang. Hal ini ditunjukan oleh nilai parameter statistic yang diperoleh tidak mengikuti pola distribusi untuk ketiga metode lainnya dan penggambaran garis teoritiknya berupa garis lengkung.



Menurut Jayadi, 2000 , Ciri Distribusi Log Person Tipe III yaitu :

1. Jika tidak menunjukkan sifat-sifat seperti ketida distribusi yaitu distribusi Gumbel, Normal maupun Log Normal
2. Garis Teoritisnya Probabilitasnya berupa garis lengkung



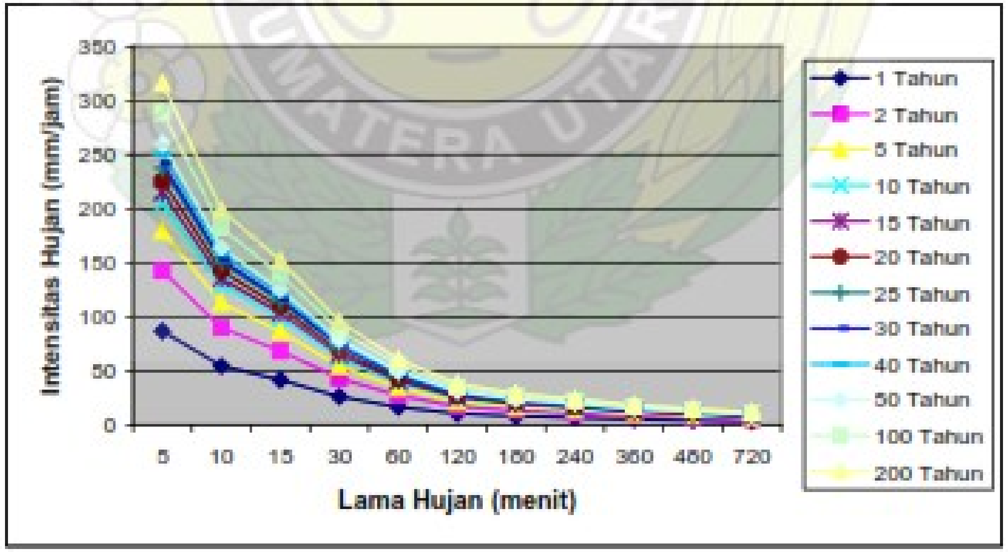
Uji Kecocokan	Nilai Tabel		Nilai Hitung
Chi-Square	3,841	>	2,0
Smirnov-Kolmogorov	0,276	>	0,1048

Nilai Intensitas Hujan

Kala Ulang (Tahun)	Hujan Rancangan (mm)
1	47.38
2	78.61
5	98.97
10	112.93
15	118.66
20	124.82
25	131.13
30	133.54
40	138.52
50	143.65
100	159.55
200	174.5

Tabel 9. Intensitas hujan jam-jaman

T	Kala Ulang											
	1	2	5	10	15	20	25	30	40	50	100	200
(menit)												
5	86.52	142.34	179.21	204.49	214.86	226.02	237.44	241.81	250.83	260.12	288.91	315.98
10	54.51	89.69	112.92	128.84	135.38	142.41	149.61	152.36	158.04	163.89	182.03	199.09
15	41.61	68.46	86.20	98.35	103.34	108.71	114.20	116.30	120.64	125.11	138.96	151.98
30	26.23	43.15	54.32	61.99	65.13	68.51	71.98	73.30	76.03	78.85	87.58	95.78
60	16.53	27.19	34.24	39.07	41.05	43.18	45.36	46.20	47.92	49.70	55.20	60.37
120	10.42	17.14	21.58	24.62	25.87	27.21	28.59	29.12	30.20	31.32	34.79	38.05
180	7.95	13.08	16.47	18.80	19.75	20.77	21.82	22.23	23.05	23.91	26.55	29.04
240	6.57	10.80	13.60	15.52	16.31	17.15	18.02	18.35	19.03	19.74	21.92	23.98
360	5.01	8.25	10.38	11.85	12.45	13.09	13.76	14.01	14.53	15.07	16.74	18.30
480	4.14	6.81	8.57	9.78	10.28	10.81	11.36	11.57	12.00	12.44	13.82	15.11
720	3.16	5.20	6.54	7.47	7.84	8.25	8.67	8.83	9.16	9.50	10.55	11.54



Gambar 4. Kurva *IDF (Intensity Duration Frequency)*

Koefisien Run Off

Jenis penutup tanah	A (km^2)	C	C x A
Hutan Primer	2.37	0.02	0.0474
Hutan Mangrove	6.92	0.02	0.1384
Hutan Sekunder/sangat terdegradasi	5.26	0.05	0.263
Tanah Terbuka/padang rumput	0.7	0.2	0.14
Kebun Campuran	387.73	0.2	38.773
Kelapa Sawit	8.52	0.4	3.408
Sawah Irigasi	3.23	0.15	0.4845
Daerah Perkotaan	15.46	0.9	9.276
Total	430.19	1.64	95.9413
Nilai C	**0,2230**		

Berdasarkan perhitungan di atas dapat dinyatakan bahwa pada kala ulang 1 tahun selama durasi hujan (waktu konsentrasi) 10,60 jam dengan intensitas hujan 3,5 mm/jam seluas 439,37 km^2 maka debit puncak yang diperoleh pada DAS Belawan sebesar 95,27 m^3/detik. Debit puncak yang diperoleh dapat dijadikan sebagai bahan dasar untuk perencanaan bangunan pengendali banjir, dimana dibangun suatu bangunan pengendali banjir yang dapat menampung debit puncak suatu aliran air sehingga dapat menghemat biaya dan waktu dalam pelaksanaan proyek pembangunan.

Tabel 11. Debit puncak di DAS Belawan

Kala Ulang (tahun)	Intensitas (mm/jam)	Debit Puncak (m^3/detik)
1	3.5	95.27
2	5.76	156.78
5	7.25	197.34
10	8.28	225.37
15	8.69	236.53
20	9.15	249.05
25	9.61	261.57
30	9.79	266.47
40	10.15	276.27
50	10.53	286.61
100	11.69	318.19
200	12.79	348.13

Data kondisi DAS Belawan yang diperoleh dari Dinas Pengairan Kota Medan dan Yayasan Leuser Indonesia adalah sebagai :

Luas total daerah pengaliran sungai Belawan (A) = 439,37 km^2

Panjang sungai Belawan = 65 km

Kemiringan = 0,00798 m/m

Tabel 3. Data penggunaan lahan pada DAS Belawan

Jenis penutup tanah	A (km^2)
Hutan Primer	2.37
Hutan Mangrove	6.92
Hutan Sekunder	5.26
Tanah Terbuka/padang rumput	0.70
Kebun Campuran	387.73
Kelapa Sawit	8.52
Sawah Irigasi	3.23
Daerah Perkotaan	15.46
Total	430.19

Sumber : Yayasan Leuser Indonesia

Tabel 4. Data curah hujan harian maksimum tahunan periode 1985-2006

No	Rmax
1	45
2	60
3	64
4	67
5	68
6	68
7	71
8	71
9	71
10	72
11	75

Koefisien Skewness	$\mathrm{Cs} = \dfrac{n\sum_{i=1}^{n}\left(X_i-\overline{X}\right)^3}{(n-1)(n-2)s^3}$
Koefisien Kurtosis	$\mathrm{Ck} = \dfrac{n^2\sum_{i=1}^{n}\left(X_i-\overline{X}\right)^4}{(n-1)(n-2)(n-3)s^4}$

Sumber: Singh, 1992.

21	118

IMPLEMENTASI METODE RASIONAL

Analisa Dimensi Saluran

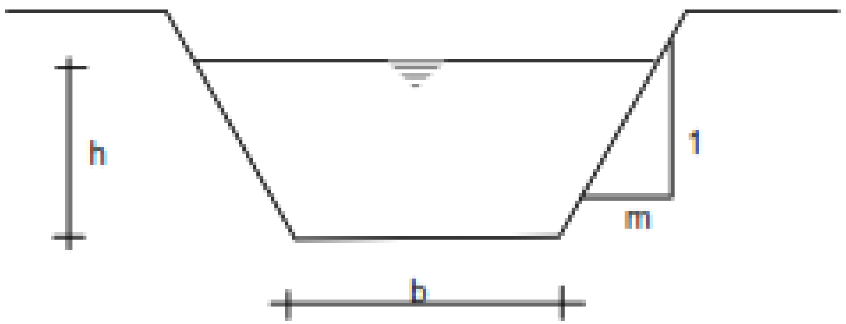
Muhamad Komarudin

PENAMPANG SALURAN EKONOMIS

Perencanaan saluran jaringan irigasi dan jaringan drainase yang pada dasarnya merupakan perencanaan penampang saluran terbuka yang mampu mengalirkan debit dari suatu lokasi ke lokasi lain dengan lancar, aman dan dengan biaya yang memadai. Dalam hal ini aspek-aspek yang perlu dipertimbangkan antara lain adalah :

- Aspek Hidrologis (Aspek Utama)
- Aspek ekonomi,
- Aspek keamanan lingkungan, dan
- Aspek estetika.

1. Saluran Bentuk Trapesium



Gambar 8 Saluran bentuk trapesium

Rumus yang digunakan :

$$A_e = (b + m.h)h$$

$$P = b + 2h\sqrt{(1 + m^2)}$$

$$R = \frac{A_e}{P}$$

Dimana :

B = lebar saluran (m)

h = dalamnya air (m)

m = perbandingan kemiringan talud

R = jari – jari hidrolis (m)

P = Keliling basah saluran (m)

A_e = Luas Penampang basah (m^2)

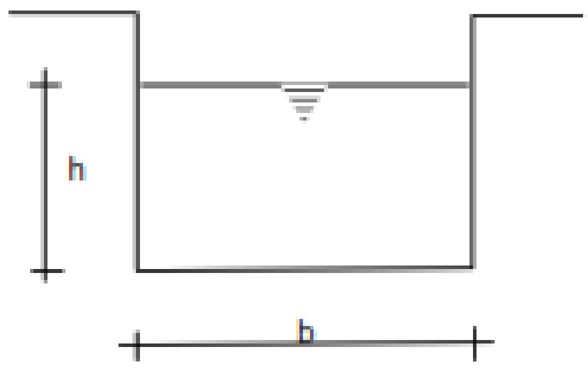
2. Saluran Bentuk Segi Empat

Rumus yang digunakan :

$$A_e = b.h$$

$$R = \frac{A_e}{P}$$

$$P = b + 2h$$



Gambar 9 Saluran bentuk segiempat

Dimana :

B = lebar saluran (m)

h = dalamnya air (m)

R = jari – jari hidrolis (m)

P = Keliling basah saluran (m)

A_e = Luas Penampang basah (m^2)

A.7.2 Penampang basah berdasarkan debit air (Q) dan kecepatan (V)

Dimensi saluran diperhitungkan dengan rumus Manning sebagai berikut :

$$Q = V.A$$

$$V = \frac{1}{n}(R)^{2/3}(i)^{1/2}$$

Dimana :

Q	:	Debit air di saluran (m^3/det)
V	:	Kecepatan air dalam saluran (m/det)
n	:	Koefisien kekasaran dinding.
R	:	Jari-jari hidraulik (meter)
i	:	Kemiringan dasar saluran
A	:	Luas penampang basah (m^2)

Tabel 20 Koefisien kekasaran dinding (n)

Tipe saluran	n
Lapisan beton	0,017 – 0,029
Pasangan batukali diplester	0,020 – 0,025
Saluran dari alam	0,025 – 0,045

A.7.3 Kemiringan Talud.

1. **Kemiringan Talud Saluran Tanah.**

   Kemiringan talud disesuaikan dengan karakteristik tanah setempat yang pada umumnya berkisar antara 1 : 1,5 s/d 1 : 4.

Tabel 21 Kemiringan Talud Bahan dari Tanah

Bahan Tanah	Kemiringan Talud (m = H/V)
Batu	0,25
Lempung kenyal, geluh	1 - 2
Lempung pasir, tanah kohesi f	1,5 - 2,5
Pasir lanauan	2 - 5
Gambut kenyal	1 - 2
Gambut lunak	3 - 4
Tanah dipadatkan dengan baik	1 - 1,5

2. **Kemiringan Talud Saluran Pasangan.**

Tabel 22 Kemiringan Talud Bahan dari Pasangan

Tinggi Air	m
h < 0,40 m	0 (dinding tegak vertikal)
0,75 > h > 0,40 m	0,25 - 0,5
H > 0,75 m	0,50 - 1,0

A.7.4 Tinggi Jagaan (F).

Tinggi jagaan minimum untuk saluran dengan pasangan direncanakan = 0,50m. Untuk saluran tanpa pasangan dengan debit tinggi jagaan sebagai berikut :

Tabel 23 Tinggi jagaan

Q	F (m)	Polder (m)
Q < 5 m^3/det	0,20 – 0,30	0,75 – 1,00
10 m^3/det > Q > 5 m^3/det	0,30 – 0,50	1,00 – 1,25
Q > 10 m^3/det	0,70 – 1,00	1,25 – 1,50

A.7.5 Kemiringan Tanah

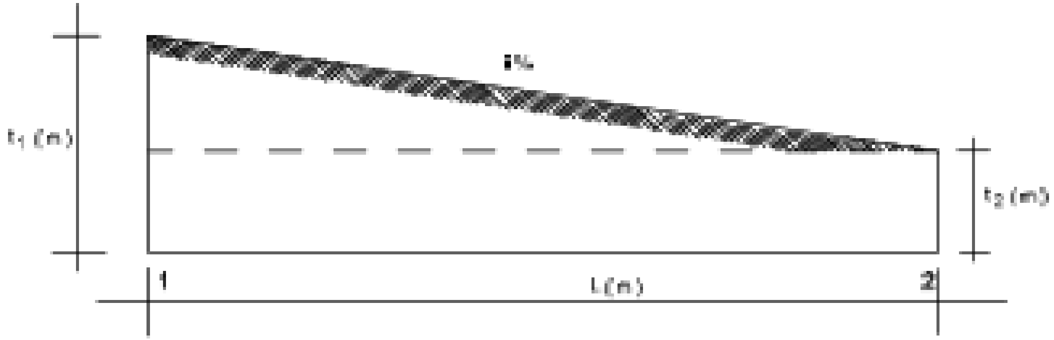
Kemiringan tanah di tempat dibuatnya fasilitas saluran drainase ditentukan dari hasil pengukuran di lapangan, dihitung dengan rumus :

$$i = \frac{t_1 - t_2}{L} x\, 100\%$$

Keterangan :

t_1 = tinggi tanah di bagian tertinggi (m)

t_2 = tinggi tanah di bagian terendah (m)



Gambar 10 Kemiringan tanah

Tabel 24 Harga n untuk rumus Manning

No	Tipe Saluran	Baik sekali	Baik	Sedang	Jelek
	SALURAN BUATAN				
1	saluran tanah, lurus teratur	0.017	0.02	0.023	0.025
2	saluran tanah yang dibuat dengan excavator	0.023	0.028	0.03	0.04
3	saluran pada dinding batuan, lurus, teratur	0.02	0.03	0.033	0.035
4	saluran pada dinding batuan, tidak lurus, tidak teratur	0.035	0.04	0.045	0.045
5	saluran batuan yang diledakkan, ada tumbuh-tumbuhan	0.025	0.03	0.035	0.04
6	dasar saluran dari tanah, sisi saluran berbatu	0.028	0.03	0.033	0.035
7	saluran lengkung, dengan kecepatan aliran rendah	0.02	0.025	0.028	0.03
	SALURAN ALAM				
8	Bersih, lurus tidak berpasir, tidak berlubang	0.025	0.028	0.03	0.033
9	seperti no.8, tetapi tidak ada timbunan atau kerikil	0.03	0.033	0.035	0.04
10	Melengkung bersih, berlubang dan berdinding pasir	0.033	0.035	0.04	0.045

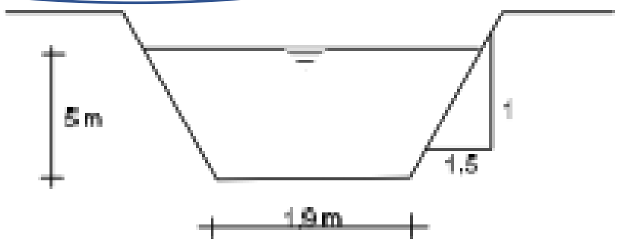
Debit air yang masuk :

$$Q_{in} = 0{,}278 C \times C_s \times I \times A$$
$$= 0{,}278 \times 0{,}73 \times 0{,}70 \times 59 \times 5$$
$$= 42 m^3 / \text{det}$$

Contoh Perhitungan 7 :

Analisa dimensi saluran trapesium dengan menggunakan data perencanaan sebagai berikut

- Debit air yang masuk (Q_{in}) = 42 m^3/det (diambil dari contoh perhitungan 5)
- Lebar saluran (b) = 5 m
- Dalamnya air (h) = 1,9 m
- Perbandingan kemiringan talud (m) = 1,5
- Kemiringan saluran yang diijinkan (i) = 0,0025
- Koefisien kekasaran Manning (n) = 0,020



Gambar 11 Kemiringan tanah

Penyelesaian :

1) Luas penampang basah saluran :

$$A_s = (b + m.h)h$$
$$= (5{,}0 + (1{,}5 x 1{,}9)) x 1{,}9$$
$$= 14{,}92 m^2$$

2) Keliling basah saluran :

$$P = b + 2h\sqrt{(1 + m^2)}$$
$$= 5 + 2(1{,}9)\sqrt{(1 + (1{,}5)^2)}$$
$$= 11{,}9 m$$

3) Jari-jari hidrolis :

$$R = \frac{A_s}{P}$$
$$= \frac{14{,}92}{11{,}9}$$
$$= 1{,}26 m$$

4) Kecepatan aliran :

$$V = \frac{1}{n} (R)^{2/3} (i)^{1/2}$$
$$= \frac{1}{0{,}020} (1{,}26)^{2/3} (0{,}0025)^{1/2}$$
$$= 2{,}91 \ m / \text{det}$$

5) Debit air yang keluar :

$$Q_{out} = V.A$$
$$= 2{,}91 \times 14{,}92$$
$$= 43{,}47 m^3 / \text{det}$$

6) Check :

$$R_{cm} = \frac{Q_{in}}{Q_{out}}$$
$$= \frac{42}{43{,}47}$$
$$= 0{,}97 \qquad (OK)$$

Latihan Soal

Mata Kuliah : Hidrologi
Dosen : 1. Ir. Feizal Manaf M.Sc.
2. Muhamad Komarudin S.Si., M.Si.
Tanggal : 14 Juni 2022
Waktu : 19.00 – 20.45 WIB

Perhatikan baik baik !
Bacalah soal dibawah ini dengan cermat dan jawablah dengan tepat dan benar. Jawaban ditulis tangan diatas kerta kosong. Kerjakan secara berkelompok dan hasil jawabannya dikirimkan ke email mkomarudin@istn.ac.id sesuai dengan batas waktu ujian yang ditetapkan.

Sebagai satu contoh kasus ada wilayah perencanaan berada di daerah perumahan di Jakarta. Wilayah ini mengalami banjir dan genangan setiap tahunnya. Penyebabnya adalah elevasi muka air banjir di sungai lebih tinggi dari elevasi tanah di daerah perumahan. Permasalahan ini diselesaikan dengan merencanakan sistem polder. Data perencanaan yang digunakan sebagai berikut :

- Luas catchment area (A) = 500 Ha
- Panjang saluran (L) = 5400 m
- Data curah hujan harian maksimum selama 20 tahun (1986 s/d 2005).
  Berikut adalah tabel data curah hujan harian maksimum selama 20 tahun (1986 s/d 2005) yang diperoleh di Stasion A (St. A). Diasumsikan Stasion A sebagai stasion curah hujan yang terdekat dengan lokasi perencanaan sistem drainase.

Tabel 1 Data curah hujan harian maksimum ($CHHm_{ax}$) St. A

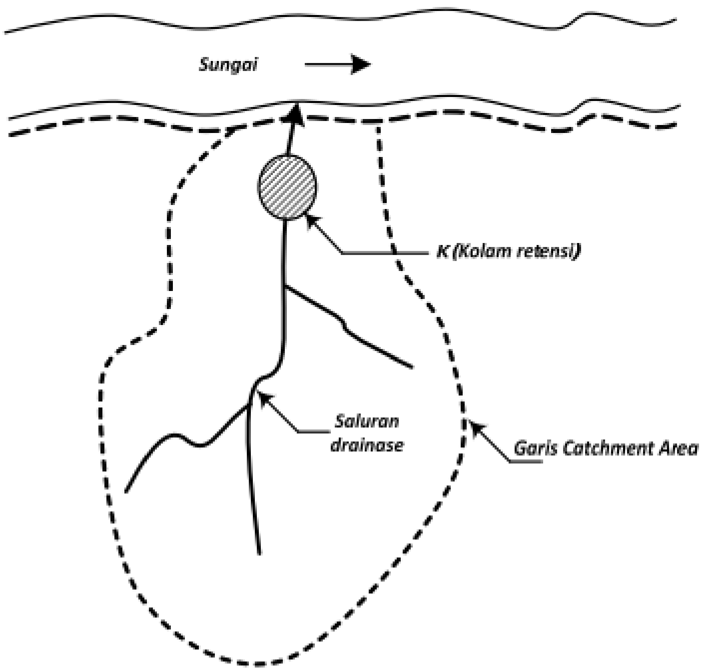
Tahun	CHH_{max}) (mm/hari)
1986	152
1987	80
1988	92
1989	130
1990	70
1991	26
1992	92
1993	79
1994	79
1995	23
1996	71
1997	112
1998	150
1999	129
2000	67
2001	92
2002	58
2003	90
2004	74
2005	87

Untuk memenuhi perhitungan hidrologi dan hidrolika perlu adanya asumsi batasan-batasan, bilamana asumsi ini terpenuhi maka analisa bisa dilaksanakan sehingga dapat dicapai sasaran penanggulangan banjir dan genangan. Asumsi perhitungan yang digunakan sebagai berikut :

1. Total Inflow – Total out flow = Storage penampungan pada waktu (t)
2. Bentuk hidrograf aliran masuk (inflow) yang digunakan sesuai bagi penggunaan rumus modifikasi Rational.
3. Rate dari flow dianggap konstan

Tugas anda melakukan perhitungan hidrologi dan hidrolika untuk perencanaan kolam retensi dan polder. Untuk itu perlu dilakukan langkah-langkah perhitungan sebagai berikut :

1. Tentukan kala ulang rencana untuk saluran di daerah Jakarta dengan luas catchment area seluas 500 Ha
    a. Dengan menggunakan data curah hujan maksimum selama 20 tahun yang terdapat pada tabel 1, analisa frekuensi hujan dengan menggunakan metode Gumbel.
    b. Dengan menggunakan data curah hujan harian maksimum selama 20 tahun yang diperoleh di tabel 1, analisa frekuensi hujan dengan menggunakan metode Log Pearson Type III.
    c. Dengan menggunakan hasil rata-rata dari metode Log Pearson III dan metode Gumbel, analisa intensitas hujan dengan berbagai kala ulang.
2. Buatlah Analisa debit banjir saluran drainase hujan periode ulang 10 tahunan dengan data perencanaan sebagai berikut :
    a. Luas catchment area (A) = 500 Ha = 5 km2
    b. Koefisien pengaliran (C) = 0,73
    c. Waktu awal (t0) = 10 menit
    d. Waktu konsentrasi (tc) = 70 menit
    e. Panjang saluran (L) = 5400 m
    f. Kecepatan rata-rata/velocity (V) = 1,5 m/det
    g. Hujan rencana kala ulang 10 tahunan (Rt)



Gambar 1 Skema sistem polder

------------------------------SELAMAT BEKERJA-------------------------------

DAFTAR HADIR PESERTA KULIAH MAHASISWA
GENAP - REGULER - TAHUN 2022/2023

FAK / JURUSAN	Teknik Sipil S1	HARI / TANGGAL	Kamis
MATAKULIAH	Hidrologi / 112019 / 2		
KELAS / PESERTA	K / 2	JAM KULIAH	19:00-20:40
KURIKULUM	2018		
DOSEN	1.Rahardjo Samiono, Ir. MT 2.Muhamad Komarudin, S.Si., M.Si	RUANG	

No	N I M	NAMA MAHASISWA	TANGGAL PERTEMUAN								JUMLAH
			8/6	15/6	22/6	06/7	13/7	20/7			
1	22114001	AKMAL KHOLIKUL IHKSAN	✓	✓	✓	✓	✓	✓			
2	22114002	MUHAMAD TRI REVALDI KARLITOS	✓	✓	✓	✓	✓	✓			

CATATAN : Perubahan peserta hanya diperkenankan bila ada persetujuan tertulis dari Pelaksana Jurusan.

Jakarta, ...

Dosen Pengajar,

(Rahardjo Samiono, Ir. MT)

17/03/2023

BERITA ACARA PERKULIAHAN
(PRESENTASI KEHADIRAN DOSEN)
SEMESTER GENAP TAHUN AKADEMIK 2022-2023
PROGRAM STUDI TEKNIK SIPIL S1 -FTSP-ISTN P2K

Mata Kuliah	**: Hidrologi**	**Semester**	**: 2**
Dosen	**: Ir. Rahardjo Samiono., MT**	**SKS**	**: 2**
	: Moh. Komarudin., S.Si, M.Si	**Kelas**	**: K**
Hari	**: Kamis**	**Ruang**	**: B.1**
Jam	**: 19:00 - 20:40**		

NO	TANGGAL	MATERI KULIAH	JML MHS HADIR	TANDA TANGAN DOSEN
1	30/3	Pengantar Hidrologi	1	
2	6/4.	Hujan (presipitasi	2	
3	13/4.	Evaporasi & infiltrasi	1.	
4	27/4	limpasan	2	
5	4/5	latihan soal	2	
6	11/5	Membahas latihan soal.	1	
7	25/5	Membahas lat soal	2	
8		UJIAN TENGAH SEMESTER (UTS)		

Dosen Mengajar

(....)

BERITA ACARA PERKULIAHAN
(PRESENTASI KEHADIRAN DOSEN)
SEMESTER GENAP TAHUN AKADEMIK 2022-2023
PROGRAM STUDI TEKNIK SIPIL S1 -FTSP-ISTN P2K

Mata Kuliah	**: Hidrologi**	**Semester**	**: 2**
Dosen	**: Ir. Rahardjo Samiono., MT**	**SKS**	**: 2**
	: Moh. Komarudin., S.Si, M.Si	**Kelas**	**: K**
Hari	**: Kamis**	**Ruang**	**: B.1**
Jam	**: 19:00 - 20:40**		

NO	TANGGAL	MATERI KULIAH	JML MHS HADIR	TANDA TANGAN DOSEN
9	8/6	Metode Rasional	2	
10	15/6	Analisa Frekwensi CH utk Rancangan Debit Banjir	2	
11	22/6	Aplikasi dan Pemecahan Masalahnya Ranc. Qp.	2	
12	06/7.	Debit Aliran Sungai	2	
13	13/7	Penampang Saluran ~~Ekonomis~~	2	
14	20/7	Penampang Saluran Majemuk	2	
15	27/7.	Latihan Soal.	2	
16		UJIAN AKHIR SEMESTER (UAS)		

Dosen Mengajar

(....)