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Evaporation from *Pinus caribaea* plantations on former grassland soils under maritime tropical conditions

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Abstract. Wet canopy and dry canopy evaporation from young and mature plantations of *Pinus caribaea* on former grassland soils under maritime tropical conditions in southwestern Viti Levu, Fiji, were determined using micrometeorological and hydrological techniques. Modeled annual evaporation totals (ET) of 1926 and 1717 mm were derived for the 6- and the 15-year-old stands, respectively. Transpiration made up 72% and 70% of annual ET, and modeled rainfall interception by the trees and litter layer was 20–22% and 8–9% in the young and the mature stands respectively. Monthly ET was related to forest leaf area index and was much higher than that for the kind of tall fire-climax *Pennisetum polystachyon* grassland replaced by the forests. Grassland reforestation resulted in a maximum decrease in annual water yield of 1180 mm on a plot basis, although it is argued that a reduction of (at least) 500–700 mm would be more realistic at the catchment scale. The impact of reforesting grassland on the water resources in southwest Viti Levu is enhanced by its location in a maritime, seasonal climate in the outer tropics, which favors a larger difference between annual forest and grassland evaporation totals than do equatorial regions.

1. Introduction

Out of concern about the potentially adverse effects on soils, hydrology, and climate of the continuing conversion of the world's tropical rain forests to other, mostly agricultural, land uses, considerable effort has gone into the evaluation and prediction of such effects during the last two decades [*Lal*, 1987; *Bruijnzeel*, 1990, 1996, 1998; *Fritsch*, 1992; *Malmer*, 1996; *Gash et al.*, 1996]. Generally, reductions in tropical forest cover of at least 30% are needed to bring about any detectable change in streamflow [*Gilmour*, 1977; *Subba Rao et al.*, 1985; *Jetten*, 1994]. While the general trend of increased water yields following the removal of a progressively larger proportion of the vegetation [*Bosch and Hewlett*, 1982] is well established for the humid tropics as well [*Bruijnzeel*, 1990], the associated range reported for the initial increase in streamflow after total forest clearance is quite large (125–820 mm yr^{-1}). This variation can be explained only partially by differences in rainfall between locations or years. Other factors affecting postclearing rates of growth and water uptake (e.g., soil fertility status, elevation), runoff disposal (catchment steepness, soil depth), or both (notably the degree of surface disturbance by machinery or fire) play a role as well [*Bruijnzeel*, 1996, 1998]. In addition, *Shuttleworth* [1989] suggested that compared to mid-continental sites, tropical deforestation is likely to have the most effect on river flow (though not necessarily on climate) at continental edge and island locations. In the absence of experimental data for tropical island conditions Shuttleworth had to base his contention on a comparison of the micrometeorology of a rain forest in central Amazonia and a spruce plantation in Wales, United Kingdom, as well as on the reported contrast in rainfall interception by tropical forests in central Amazonia [*Lloyd and Marques*, 1988] and Java [*Calder et al.*, 1986].

The repeated burning of forest (often in combination with subsequent grazing by cattle) has produced extensive degraded grassland areas in such "maritime" tropical locations as the Philippines [*Quimio*, 1996], Indonesia [*MacKinnon et al.*, 1996], and the southwestern Pacific [*Drysdale and Rawaqa*, 1987]. Given the extent of hydrologically and ecologically disrupted land in these areas, calls for massive reforestation programs have become more frequent. As a result, increasingly large areas of semiproductive grassland and scrubland are being converted to plantation forests for wood production [*Evans*, 1992]. The dry western side of Viti Levu, the largest island in the Fiji archipelago, provides a case in point. The area was deforested well over 100 years ago, and overgrazing and repeated burning resulted in the forming of a tall, fire-climax grassland where *Pennisetum polystachyon*, introduced in Fiji in the 1920s, is the dominant species. Between the early 1970s and 1990, a reforestation program converted some 23,230 ha of grassland to plantation forest (*Pinus caribaea* Morelet var. *hondurensis* Barr. and Golf.). However, major reductions in the streamflow from several catchment areas used for the supply of water to the adjacent lowlands were inferred from spot measurements of minimum flows after reforestation [*Kammer and Raj*, 1979]. The resulting conflict of interests between Fiji Pine Ltd. and the Department of Public Works led to the initiation in late 1989 of a study of the micrometeorology, hydrology, and soil fertility aspects of tall fire-climax grassland and an age series of pine plantations [*Waterloo*, 1994]. The present paper

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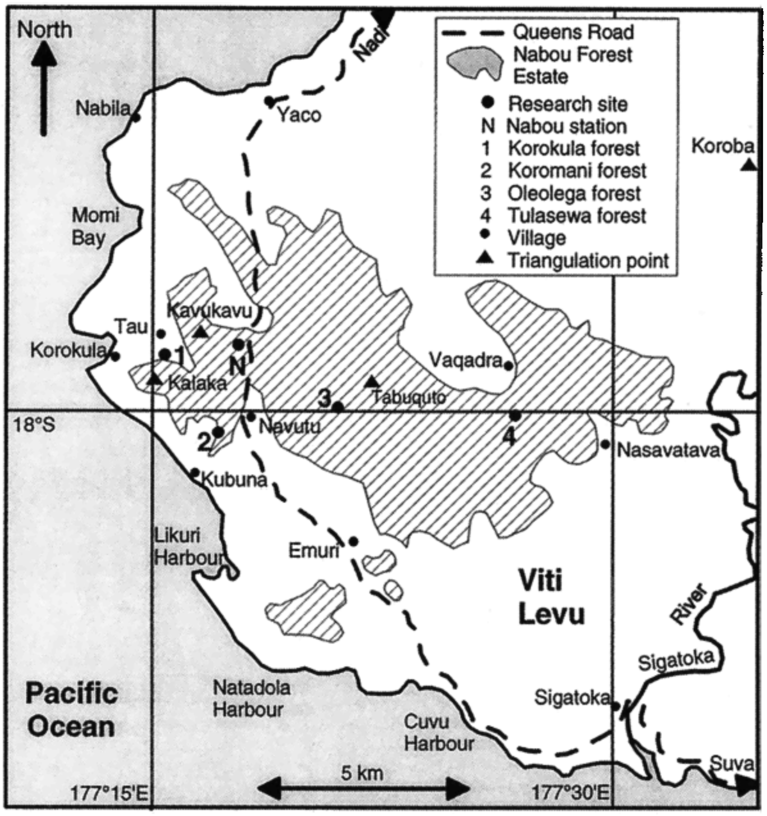


Figure 1. Location of the study sites within the Nabou Forest Estate, southwestern Viti Levu, Fiji: 2, Koromani, 15-year-old pine forest; 3, Oleolega pine-forested catchment; 4, Tulasewa, six-year-old pine forest; and N, Nabou, seasonal grassland site.

summarizes the micrometeorological, soil, and catchment hydrological evidence obtained for a young (6 years old) and a mature (15 years old) pine plantation forest during several field campaigns held between November 1989 and September 1991. A detailed discussion of the micrometeorology and water use of the grassland is presented elsewhere (M. J. Waterloo et al., Micrometeorology and water use of tall grassland in southwestern Viti Levu, Fiji, manuscript in preparation, 1999) (hereinafter referred to as Waterloo et al., manuscript in preparation, 1999).

Evidence of diminished water yields following afforestation of natural grassland and scrubland with pines or eucalypts has come from various seasonal warm-temperate and subtropical locations, notably in South Africa (reductions of 300–450 mm yr^{-1} [*Bosch*, 1979, 1982; *Van Lill et al.*, 1980; *Smith and Scott*, 1992]), India (reductions of 10–120 mm yr^{-1} [*Mathur et al.*, 1976; *Mathur and Sajwan*, 1978; *Samraj et al.*, 1988]), and southeastern Brazil (120–350 mm yr^{-1} [*Lima et al.*, 1990]). However, very few data are available in this respect for more equatorial latitudes [*Bailly et al.*, 1974; *Daño*, 1990], whereas in addition, the usefulness of these data is limited because of problems with catchment leakage, which leads to unrealistically high values for the water use of the original grasslands [*Bruijnzeel*, 1990]. *Hodnett et al.* [1996] inferred a difference in water use of about 350 mm between rain forest and man-made pasture at the end of the dry season in central Amazonia. However, if the contention of *Shuttleworth* [1989] is true, much larger differences may be expected in the case of reforesting seasonal grassland with fast-growing pine plantations under the maritime tropical climatic conditions prevailing in the Fiji archipelago. This paper is the first to report on total evaporation (ET), its component wet canopy and dry canopy evaporation rates [*Monteith*, 1965], and the effect on catchment water yield of fast-growing pine plantations on former grassland soils in the lowland humid tropics [*Bruijnzeel*, 1997].

2. Study Area

2.1. Site Description

The study was conducted in the Nabou forest estate of Fiji Pine Ltd. in the southwestern part of Viti Levu (Figure 1). Two stands of 6- and 15-year-old pine trees were selected in the undulating terrain to represent young, vigorously growing pines and more mature and slower growing trees, respectively. The young plantation was located close to the hamlet of Tulasewa (18°00'S, 177°27'E) at an elevation of 116 m above sea level (asl), whereas the older stand was located near Koromani (18°01'S and 177°18'E) at an elevation of 90 m asl (Figure 1). The pines at the Tulasewa site had been planted at a spacing of 3×3 m (equivalent to a stocking of 1111 trees ha^{-1}), and the stocking in January 1990 was 825 trees ha^{-1}. The average diameter (measured at 1.35 m), basal area, and height of the trees were 0.16 ± 0.04 m, 18.1 m^2 ha^{-1}, and 11.6 ± 2.3 m, respectively. The trees were actively growing, and the current annual increase in basal area was 3.9 m^2 ha^{-1} yr^{-1}. The leaf area index (LAI) of the pine trees was determined at 3.5 ± 0.9 m^2 m^{-2} in July 1990 using destructive sampling techniques, which were checked against results obtained from light attenuation measurements [*Pierce and Running*, 1988; *Waterloo*, 1994]. The dense undergrowth was dominated by *Pennisetum* *polystachyon* grass and bracken fern. The reddish-brown soil (developed in volcanic breccias) was classified as a Udic Haplustoll according to the USDA Soil Taxonomy system and had a silty-clay to clay texture and a weak structure. Bulk density (BD) increased from 970 ± 200 kg m^{-3} in the topsoil (0–30 cm) to 1120 ± 90 kg m^{-3} in the subsoil (120–150 cm). The water holding capacity (PAW) of the upper 80 cm of the soil, where the majority of the fine roots occurred, was estimated from soil moisture retention curves at 150 ± 53 mm ($n = 5$). The median saturated hydraulic conductivity (K_s, determined with a permeameter on undisturbed, 100-cm^3 soil cores) ranged from 1.5 m d^{-1} ($n = 12$) in the topsoil to 0.012 m d^{-1} ($n = 5$) below a depth of 75 cm. Few fine roots were observed below 90 cm in a soil pit down to a depth of 155 cm [*Waterloo*, 1994].

Although the spacing of the trees upon planting in 1975 at the Koromani site had also been 3 × 3 m, stand density had been reduced to 621 trees ha^{-1} by January 1990 as a result of repeated cyclone passage over the years. The average tree diameter, basal area, and height were 0.25 ± 0.06 m, 31.6 m^2 ha^{-1}, and 17.5 ± 2.6 m, respectively. The forest had an LAI of 4.0 ± 0.7 m^2 m^{-2}. Growth by the mature forest at Koromani was less vigorous than that of the young stand at Tulasewa, with a current annual increase in basal area of 1.7 m^2 ha^{-1} yr^{-1}. The forest was defoliated by cyclone Sina on November 27–28, 1990. Cyclone-induced needle fall amounted to about 6000 kg ha^{-1}, part of which consisted of dead needles suspended in the canopy [*Waterloo*, 1994]. This suggested that the LAI was reduced by 2.8 m^2 m^{-2} at the most and may have ranged between 1.2 and 2.0 m^2 m^{-2} shortly after the cyclone event. The forest quickly produced new foliage during the ensuing wet season and eight months later (July 1991) the LAI was determined as 3.1 ± 0.8 m^2 m^{-2}. The stand basal area was 33.5 m^2 ha^{-1} in September 1991. Undergrowth was sparse and consisted mainly of grass, scrubs, and deciduous saplings. The reddish-brown soil (developed from andesitic rock) was classified as a Typic Eutrustox with a silty-clay to clay texture and a weak structure. Values for BD increased from 990 ± 160 kg m^{-3} in the topsoil (0–10 cm, $n = 10$) to 1360 ± 50 kg m^{-3} in the subsoil (80–120 cm; $n = 6$) whereas the median K_s ranged from 3.2 m d^{-1} in the topsoil to 0.23 m d^{-1} below a depth of 30 cm. The PAW was estimated at 112 ± 47 mm for the upper 80 cm of the profile ($n = 5$). Few fine roots were observed below 80 cm in a soil pit down to a depth of 170 cm [*Waterloo*, 1994].

The characteristics of the *Pennisetum polystachyon* grassland reference site (Nabou, 17°57′S, 177°19′E, 91 m asl; Figure 1) have been described in detail by Waterloo et al. (manuscript in preparation, 1999). Summarizing, the grassland site is not managed and had not been burned for at least 5 years at the start of the study. The grass shows a pattern of active growth during the wet season followed by a gradual dying off of the aboveground parts during the dry season. This pattern was reflected in a variation in grass height (0.9–1.8 m) and a 10-fold variation in LAI (0.2–2.1 m^2 m^{-2}). The PAW value of the upper 80 cm of the clay to clayloam soil (a Udic Argiustoll) at Nabou, where most of the roots were found, was estimated at 100 ± 6 mm ($n = 2$).

2.2. Climate

The climate of Viti Levu is seasonal maritime tropical and strongly influenced by the presence of several north-south stretching mountain ranges dividing the island in a humid (rainfall >3000 mm yr^{-1}) climate at the windward, eastern side and a dry (rainfall <2000 mm yr^{-1}) climate at the leeward, western side [*Krishna*, 1980]. The present study was conducted in the dry zone. The following summary of the local climate is largely based on long-term (1942–1985) averages for Nadi Airport, which is situated some 25 km to the north of the study area [*Basher*, 1986a; *Reddy*, 1989b, c]. Average annual rainfall (P) at Nadi Airport amounts to 1867 mm (range of 864–2983 mm), against 1707 mm (range of 826–2498 mm) at the Nabou grassland station (1974–1995 [*Reddy*, 1989a]). Rainfall in the Nabou estate shows a distinct seasonal pattern with low average monthly totals (<100 mm) between May and October, and high values (up to 300 mm) between November and April. Average temperatures at Nadi Airport range from 23.5°C in July to 27.1°C in January, whereas average relative humidities range from 76% at the end of the dry season (October) to 84% at the end of the wet season (March). Average shortwave radiation totals range from 14 MJ m^{-2} d^{-1} in June to 22 MJ m^{-2} d^{-1} in December. Easterly to southeasterly trade winds blow throughout the year but are most pronounced during the dry season. Wind speeds are generally low with an average value of 2.8 ± 0.4 m s^{-1}. Wind speeds are slightly higher during the dry season than during the wet season, with monthly averages of 3.4 ± 1.4 m s^{-1} in October versus 2.4 ± 1.0 m s^{-1} in April. Strong winds are observed only during the passage of cyclones, which occur on average at a frequency of about once every 1.3 years. The long-term (1947–1985) annual Penman open water evaporation (E_0) at Nadi Airport averages 1606 mm yr^{-1}. Monthly averages range from 3.0 mm d^{-1} during the southern winter (June) to 5.5 mm d^{-1} during the southern summer (December) [*Reddy*, 1989c].

Table 1. Instrument Configuration of the Meteorological Masts at the Tulasewa Forest Site for Two Periods and at the Koromani Forest for One Period

Level	Instrument	Tulasewa*	Tulasewa†	Koromani‡
High	Wind vane	12.5	21.9	32.3
	Anemometer	12.5	21.1	32.1
	Rotronic T-RH	12.5	21.1	32.1
	Thermocouple	···	21.2	31.9
Mid	Anemometer	11.0	17.0	27.0
	Rotronic T-RH	···	17.0	27.0
	Thermocouple	···	17.0	26.9
Low	Anemometer	9.5	13.0	24.1
	Rotronic T-RH	9.5	13.0	24.1
	Thermocouple	···	13.0	24.1
	Pyranometer	11.0	12.8	21.9
	Net radiometer	11.0	12.5	21.9
	Albedometer	···	12.6	21.9
	Pluviograph	8.7	11.8	22.0
Canopy		11.6	12.6	16.5

Canopy heights added for comparison.
*November 30, 1989, through April 23, 1990.
†May 8 through November 27, 1990.
‡April 27 through September 20, 1991.

3. Instrumentation and Methods

3.1. Micrometeorological Measurements

Micrometeorological measurements were made at the Tulasewa and Koromani forest sites to determine forest evaporation. The instrument configurations of the masts at the two sites are given in Table 1. At Tulasewa, basic wet-season climatic data were collected at canopy level in a 12.5 m mast between November 30, 1989, and April 23, 1990 (145 days).

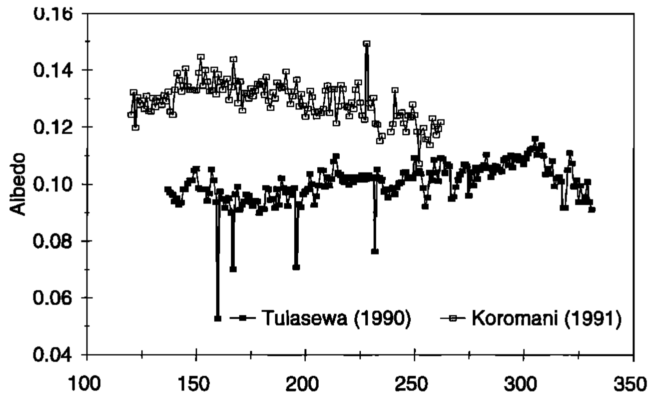


Figure 2. Seasonal variation in daily albedo of the young (Tulasewa) and the mature (Koromani) pine forest.

More intensive dry-season measurements were made well above the canopy in a 21.9-m mast between May 8 and November 27, 1990 (204 days). Similar measurements were made at Koromani forest using a 32.9-m mast between April 27 and September 20, 1991 (140 days).

Solar and net radiation were measured with a Skye SP-1110 pyranometer and a Radiation and Energy Balance Systems Q*5 net radiometer, respectively. The albedo was determined with a Kipp and Zonen CM-7 albedometer. The radiation instruments had an absolute error of <5% and were placed on a support arm extending 1.5 m from the mast in such a way as to avoid shading by the mast. The soil heat flux (G) was measured with two Middleton and Co. flux plates placed at 2 cm below the soil surface under a 7- to 15-cm-thick litter layer. Care was taken not to disturb the soil and litter layer during installation of the plates. Although the accuracy of the flux plates themselves was given as 5%, the actual error may be much larger because of spatial variations in G and possible differences between the heat capacities of the soil and flux plates. Air temperature and humidity were measured with non-aspirated Rotronic MP100F temperature and humidity probes placed in Gill radiation screens. The platinum resistance thermometer (PRT) had an accuracy better than 0.2°C and a long-term stable precision of 0.1°C. The accuracy of the humidity sensor was given as better than 1% and a long-term stable precision of 0.5% of the relative humidity. However, above values of 97% humidity very small changes in temperature could cause condensation on the sensors, and readings would not have their normal accuracy. Apparent relative humidities in excess of 100% were sometimes experienced during rainfall or at night. Whenever this occurred, values were set at 100%. Additional temperature measurements were made using fast-response chromium-constantane wire thermocouples at a frequency of 0.5 Hz to derive sensible- and latent-heat fluxes using the temperature variance energy balance (TVEB) method [*Tillman*, 1972; *De Bruin*, 1982; *Vugts et al.*, 1993]. All temperature and humidity instruments had been calibrated before installation at their respective levels above and within the canopy and followed by recalibrations during the observation period by shifting the position of one of the instruments to higher levels for several days at a time. Wind speeds were measured with Vector Instruments A101M/L anemometers having a stalling speed of 0.15 m s^{-1} and an accuracy of 1–2% of the wind speed up to values of 75 m s^{-1}. No corrections were made for overspeeding or stalling. The anemometers were positioned on 0.8-m support arms pointing toward the general wind direction (southeast). Wind direction was measured with a potentiometer-type wind vane (Vector Instruments W200P). The fin had a threshold value of 0.6 m s^{-1} and a resolution of about 1°. All instruments were connected to a Campbell 21X datalogger/multiplexer system. Thermocouple data were preprocessed at 5-min intervals to avoid any influence of trends in air temperature on half-hourly standard deviations. The other instruments were sampled at 30-s intervals. All averages and standard deviations were calculated over 30-min periods. Above-canopy rainfall (5-min totals) was measured with a Campbell Scientific ARG-100 tipping bucket rain gauge (0.2-mm tip) connected to the 21X datalogger. In addition, at least two standard rain gauges (100-cm^2 orifice at 1.3 m) were placed in nearby clearings making sure that no obstructions were present above a maximum angle of 30°.

3.2. Forest Hydrological Measurements

Throughfall was measured with 20 standard rain gauges (100-cm^2 orifice) per site that were regularly relocated in a random manner [*Lloyd and Marques*, 1988]. Measurements at Tulasewa were made from November 30, 1989, until August 30, 1991, whereas those at Koromani covered the period from July 23, 1990, until October 2, 1991. As such, both prehurricane and posthurricane throughfall data were collected. Stem flow was measured from October 1990 until August 1991 on five trees in the Tulasewa stand. No measurements were made at Koromani. Spiral-type PVC gutters were connected to 4.1-L containers by plastic tubing. Because of the limited capacity of the containers, reliable measurements were obtained only for rainfall events <16 mm. The throughfall and stem flow data were used to derive values for the canopy saturation value (S), free throughfall coefficient (p), trunk storage capacity (S_t), and the fraction diverted to the tree trunks (p_t) for later use in the analytical model of rainfall interception of *Gash*, [1979].

Measurements of soil moisture were made at least once a week with a capacitance probe (Didcot Instruments Ltd.) at 2-cm intervals in five access tubes per site down to a maximum depth of 1.2 m. Daily values of soil moisture deficit (SMD) in the upper 70 cm of the soil were calculated from rainfall, the Penman open water evaporation, and the water storage at field capacity using an empirical water balance model similar to that of *Calder et al.* [1983] and calibrated against measured deficits. Because the depth of the root network exceeded that of the soil water depletion technique [*Cooper*, 1979]. Instead, they were used in modeling the dependence of the canopy surface resistance (r_s) on soil water status by using the Penman-Monteith equation [*Monteith*, 1965] inversely (see below).

4. Results

4.1. Albedo, Net Radiation, and Soil Heat Flux

The shortwave reflection coefficient of the pine trees at Tulasewa was measured from mid-May until late November 1990, when cyclone Sina effectively destroyed the stand. The average albedo was 0.100 ± 0.007 (n = 195). The average reflection coefficient as measured between late April and mid-September 1991 at Koromani was significantly higher (significance level α = 0.01, student's t test) at 0.126 ± 0.008 (n = 123). The variation in daily albedo and the average diurnal patterns at the two sites are given in Figures 2 and 3, respectively. The diurnal course of the albedo was calculated from

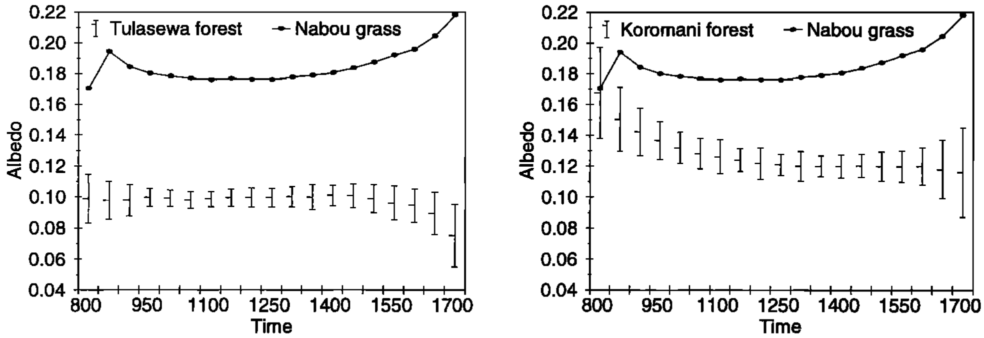


Figure 3. Average diurnal patterns of the dry season albedo of young (Tulasewa, May 17 through September 19, 1990) and mature (Koromani, May 17 through September 19, 1991) pine forests. The wet season average (January 31 through April 9, 1991) for *Pennisetum polystachyon* grassland (Waterloo et al., manuscript in preparation, 1999) added for comparison.

30-min daytime averages with global radiation inputs in excess of 100 W m^{-2}. Shortwave reflection by the young pines was markedly reduced (down to values as low as 0.05) when the canopy was thoroughly wetted by rain and radiation conditions were low. However, this phenomenon was hardly observed for the mature stand (Figure 2). The daily data for the forest at Tulasewa suggest a small seasonal variation in albedo with a significantly lower value ($\alpha = 0.01$) at the end of the wet season (0.095 ± 0.010, $n = 30$) than at the end of the dry season (0.106 ± 0.003, $n = 30$). By contrast, the albedo of the forest at Koromani decreased significantly ($\alpha = 0.01$) from 0.131 ± 0.004 ($n = 30$) to 0.121 ± 0.005 ($n = 30$) during the dry season. The albedo of the pines at Tulasewa was remarkably constant throughout the day, only to decline after 1600 LT. Conversely, the reflection coefficient of the mature pines showed only a slight decrease during late afternoon, but values were clearly above afternoon levels during the first half of the morning (Figure 3).

Linear regression analysis was applied to half-hourly (W m^{-2}) and daily (MJ m^{-2}) data pairs of global radiation (R_g) and net radiation (R_n). The resulting expressions for half-hourly and daily totals for dry season conditions (May 16 through September 19, 1990) for the young pine stand at Tulasewa are given by (1) and (2), respectively. The overall expression for daily totals (November 30, 1989, through November 27, 1990) is given in (3).

$$R_n = -21.27(\pm 13.17) + 0.867(\pm 0.001) \cdot R_g \qquad (1)$$
$$r^2 = 0.99, \; n = 6096$$

$$R_n = -0.19(\pm 0.53) + 0.757(\pm 0.011) \cdot R_g \qquad (2)$$
$$r^2 = 0.97, \; n = 127$$

$$R_n = -0.07(\pm 0.86) + 0.764(\pm 0.008) \cdot R_g \qquad (3)$$
$$r^2 = 0.97, \; n = 340$$

The corresponding regression equations for half-hourly (4) and daily (5) data pairs for the mature plantation at Koromani (May 16 through September 19, 1991) were

$$R_n = -38.6(\pm 21.4) + 0.832(\pm 0.001) \cdot R_g \qquad (4)$$
$$r^2 = 0.99, \; n = 6090$$

$$R_n = -0.58(\pm 0.92) + 0.638(\pm 0.017) \cdot R_g \qquad (5)$$
$$r^2 = 0.92, \; n = 122$$

Statistical tests using the dummy variable technique of *Kleinbaum and Kupper* [1978] indicated that at Tulasewa the dry season expression (2) coincided with the overall expression (3), suggesting no significant seasonal variation in the relation between R_g and R_n. The dry season expressions (equations (2) and (5)) for the two sites were not coincident or parallel ($\alpha = 0.01$), indicating that R_n at Tulasewa was significantly higher than at Koromani for similar R_g conditions.

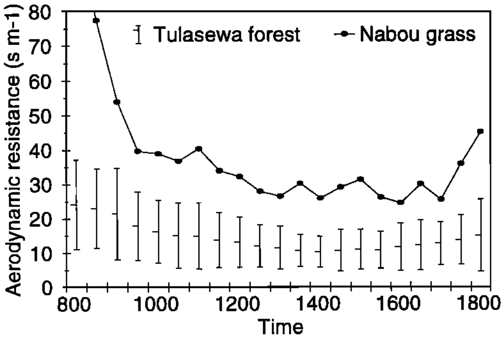
The soil heat flux (G) below the young pines varied between -5 ± 2 and 15 ± 8 W m^{-2} over a 145-day period that covered the end of the wet season plus the entire dry season. The diurnal range was smaller in the mature Koromani forest, where G varied between -3 ± 1 and 1 ± 2 W m^{-2}. Daily totals of G ranged from -0.4 to 0.2 MJ m^{-2} d^{-1} at Tulasewa, whereas values at Koromani were negative throughout the 1991 observation period, varying between -0.2 and -0.1 MJ m^{-2} d^{-1}. In both cases the soil heat fluxes were very small (usually less than 1%) compared to incoming shortwave radiation totals. The latter averaged 17.5 ± 5.9 MJ m^{-2} d^{-1} at Tulasewa and 14.6 ± 4.8 MJ m^{-2} d^{-1} at Koromani. As such, the soil heat flux term may be safely ignored in the forest energy balance. Because the same holds for the biophysical energy storage terms [*Waterloo*, 1994] the energy that is effectively available for partitioning between the sensible-heat flux (H) and the latent-heat flux (λE) can be approximated by R_n (in the absence of advected energy).

4.2. Aerodynamic Resistance

The aerodynamic resistance of the pine stands to evaporation (r_a, in s m^{-1}) was calculated from above-canopy wind speed (u_z, in m s^{-1}) measured at the highest level (z) above the canopy using the following equation [*Thom*, 1975]:

$$r_a = \frac{\left(\ln \dfrac{z-d}{z_0}\right)^2}{k^2 \cdot u_z} \qquad (6)$$

where d is the displacement length and z_0 the roughness length of the forest (both in meters) and k is Von Kármán's constant (0.4). The surface roughness parameters d and z_0 depend on the height, density, and flexibility of the vegetation. They were determined from measured above-canopy wind speed profiles using the "graphical method" of *Thom* [1975]. The wind speed profile data consisted of half-hourly measurements collected under near-neutral atmospheric conditions with the wind



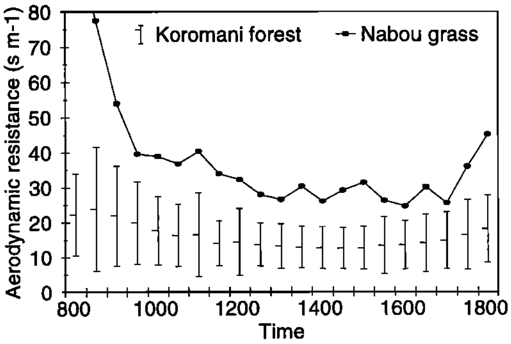


Figure 4. Average daytime patterns of the aerodynamic resistance of the young (Tulasewa, May 8 through November 27, 1990) and the mature (Koromani, April 29 through September 19, 1991) pine forest. Average for *Pennisetum polystachyon* grassland (March 23 through April 6, 1995) (Waterloo et al., manuscript in preparation, 1999) added for comparison.

speed at the lowest level in excess of 1 m s^{-1} and with adequate fetch conditions in the prevailing wind direction. These conditions were satisfied during 971 and 244 half-hourly periods at Tulasewa and Koromani, respectively.

The graphical method gave values of 8.6 m and 11.9 m for $z_0 + d$ at Tulasewa and Koromani, respectively [*Waterloo*, 1994]. Rather similar z_0 values were obtained for the two forests, namely, 1.5 m (Tulasewa) and 1.4 m (Koromani). However, because of the difference in stand height (h_v) between the two sites (12.5 m at Tulasewa and 17.5 m at Koromani), the ratio z_0 to h_v varied from 0.12 (Tulasewa) to 0.08 (Koromani). The resulting displacement lengths of the two forests did reflect the contrast in forest stature, d being 7.1 m for the young stand and 10.4 m for the mature stand. The ratios of d to h_v, however, were again similar at 0.56 (Tulasewa forest) and 0.58 (Koromani forest). Aerodynamic resistances were then calculated from wind speeds using (6). This resulted in the following expressions for the forests at Tulasewa (7) and Koromani (8):

$$r_a = \frac{31.2}{u_{21.1}} \tag{7}$$

$$r_a = \frac{47.0}{u_{32.1}} \tag{8}$$

Daily dry season (May 8 to September 19) wind speeds above the inland forest near Tulasewa (z = 21.1 m) ranged from 1.2 to 4.6 m s^{-1} (average 2.2 ± 0.8 m s^{-1}, n = 135 days) versus 1.6 to 6.1 m s^{-1} (average 3.0 ± 0.8 m s^{-1}, n = 132 days) at the Koromani site (z = 32.1 m) which was located closer to the ocean (Figure 1).

The average daytime patterns of r_a for the two forests were very similar from about 1130 LT onwards, after which r_a remained close to 10 s m^{-1} (Figure 4). Values of r_a for both forests reached a maximum in the early morning, with the peak for the older forest being both more pronounced and occurring slightly later (Figure 4).

4.3. Modeling Annual Forest Evaporation

4.3.1. Derivation of surface resistances. Under nonadvective conditions the simplified energy balance of a vegetated surface may be written as

$$R_n = H + \lambda E \tag{9}$$

where the terms are as defined before (all in W m^{-2}). *Tillman* [1972] showed that the standard deviation (σ_T) of high-frequency temperature (T) measurements could be related to the sensible-heat flux H according to

$$H = h_\sigma \rho c_p \sqrt{(z-d)\frac{g}{T}\cdot \sigma_T^{1.5}} \tag{10}$$

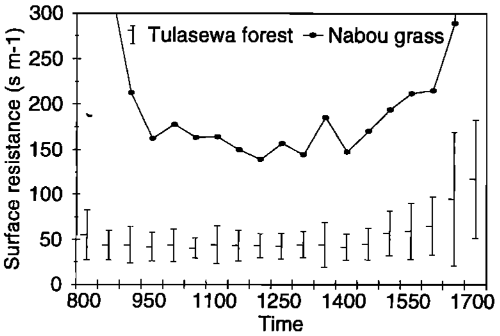
where h_σ is a constant (0.7) [*Wijngaard and Cote*, 1971], ρ is the density of air (kg m^{-3}), c_p is the specific heat of air at constant pressure (J kg^{-1} K^{-1}), z is the thermocouple height (meters), d is the displacement length of the forest (meters), and g is the acceleration due to gravity (m s^{-2}). The temperature variance energy balance method (TVEB) combines the energy balance equation (9) with the sensible-heat flux equation (10) to derive estimates of λE under dry canopy conditions [*De Bruin*, 1982; *Vugts et al.*, 1993]. The calculated λE values include the energy used for evaporation from a dry canopy (transpiration, E_t) and evaporation of moisture stored in the litter layer during dry conditions (E_l). Because the soils of the study sites were covered with a thick layer of grass and needle litter, evaporation from the soil was assumed to be negligible at all times.

The TVEB method was applied to obtain half-hourly values of H and λE from net radiation and thermocouple temperature data collected above the respective forest canopies. Because the method can only be applied usefully under unstable atmospheric conditions (i.e., Richardson number $Ri < -0.1$) [*Tillman*, 1972; *Thom*, 1975; *Frumau*, 1993], which in the study area generally prevailed between 0800 and 1700 LT on dry days [*Waterloo*, 1994], calculations were restricted to daytime conditions (i.e., $R_n > 0$ W m^{-2}) and a dry canopy. Nighttime evapotranspiration was assumed to be negligible. Furthermore, fetch requirements in the direction of the prevailing wind should be satisfied whereas soil water deficit data should be available as well. This resulted in the selection of 615 and 687 half-hour periods for the stands at Tulasewa and Koromani, respectively.

Half-hourly evaporation rates for dry canopy conditions (E_d, i.e., $E_t + E_l$) were calculated from (9) and (10). These were then used to calculate corresponding surface resistances (r_s, in s m^{-1}) using the inverse Penman-Monteith equation [*Monteith*, 1965]:

$$r_s = \frac{\rho c_p}{\gamma}\frac{\delta e}{\lambda E} + r_a\left(\frac{\Delta(R_n - G)}{\gamma \lambda E} - \frac{\Delta}{\gamma} - 1\right) \tag{11}$$

where γ is the psychrometric constant (mbar K^{-1}), δe is the vapour pressure deficit (mbar), λ is the latent heat of vapor-



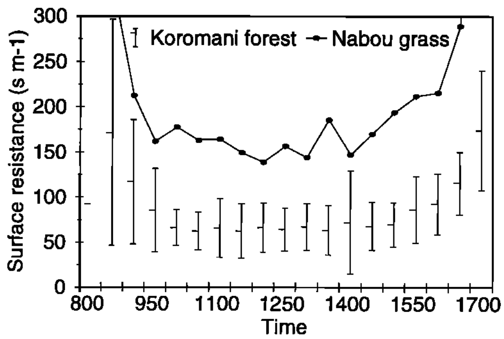


Figure 5. Average daytime patterns of the surface resistance of the young (Tulasewa) and the mature (Koromani) pine forest. Values for *Pennisetum polystachyon* grassland (Waterloo et al., manuscript in preparation, 1999) added for comparison.

ization of water (J kg^{-1}), Δ is the change of the saturation vapor pressure with temperature (mbar K^{-1}) and the available energy for evaporation approximated by R_n (W m^{-2}).

The average diurnal patterns of half-hourly values of r_s for the two forests calculated with (11) are shown in Figure 5. Values of r_s were high in the early morning (particularly at Koromani) but decreased rapidly to rather low values, which remained more or less constant until late afternoon. Typical daytime averages were 35–50 s m^{-1} for the young forest at Tulasewa to 60–80 s m^{-1} for the mature forest at Koromani (Figure 5).

Regression analysis was subsequently used to relate half-hourly values of $\ln(r_s)$ to corresponding values of $\ln(R_n)$, δe, r_a and SMD. The following best-fit relationships were obtained for the stands at Tulasewa (11) and Koromani (12), respectively:

$$r_s = \exp\,[7.838 - 0.864 \cdot \ln\,(R_n) + 0.104 \cdot \delta e + 0.005 \cdot r_a + 0.001 \cdot \mathrm{SMD}] \qquad (12)$$

$$r_s = \exp\,[8.540 + 0.949 \cdot \ln\,(R_n) + 0.086 \cdot \delta e + 0.011 \cdot r_a + 0.003 \cdot \mathrm{SMD}] \qquad (13)$$

The correlation coefficients for the regression equation relating $\ln(r_s)$ to $\ln(R_n)$ was 0.26 for the forest at Tulasewa and 0.43 for that at Koromani. Inclusion of δe increased the correlation coefficients to 0.70 and 0.74, respectively, whereas inclusion of r_a and SMD had only a minor effect, raising the respective correlation coefficients to 0.71 and 0.77.

4.3.2. Evaporation during dry canopy conditions. To extend the time for which instantaneous half-hourly rates of E_d could be determined with the TVEB method to a full year (November 30, 1989, to November 29, 1990), use was made of the Penman-Monteith equation using the forest specific relations for R_n (equations (1) and (4)), r_a (equations (7) and (8)), and r_s (equations (12) and (13)) derived in the preceding sections. A continuous record of above-canopy climatic conditions was available for the forest at Tulasewa for 349 days between November 30, 1989, and November 27, 1990. The same climatic data were used for both forests to enable a more direct comparison to be made. Daily values of R_n, T, u, RH, P, SMD, and E_0 for Tulasewa forest are shown in Figures 6a–6c.

Half-hourly periods during which the canopy was wet were excluded from the calculations (rainfall interception losses will be discussed separately below). Evaporation on dry days for which above-canopy climatic data were missing (April 24 through May 7 and November 28 and 29, 1990) was approximated by multiplying the corresponding Penman open water evaporation totals (E_0 [*Penman*, 1956]) using a modified radiation term to represent local conditions [*Waterloo*, 1994] for the nearby Nabou station times the average E_d/E_0 ratio on dry days for the preceding 30 days. The latter averaged 1.11 ± 0.07 and 0.89 ± 0.15 for the forest at Tulasewa and 0.98 ± 0.06 and 0.79 ± 0.13 for the forest at Koromani for the April/May and November periods, respectively.

Total E_d over the 365-day period amounted to 1549 mm (4.24 mm d^{-1}) at Tulasewa and 1351 mm (3.70 mm d^{-1}) at Koromani. As was to be expected, these overall averages of E_d are slightly lower than the ones obtained with the TVEB method for a limited number of selected days with unstable atmospheric conditions (i.e., 4.6 mm d^{-1} at Tulasewa versus 3.9 mm d^{-1} at Koromani). The corresponding average E_d/E_0 ratios (dry days only) were 1.02 ± 0.13 (range 0.55–1.31) for the 6-year-old forest and 0.89 ± 0.12 (range 0.42–1.13) for the 15-year-old forest.

4.3.3. Rainfall interception. Evaporation during wet canopy conditions was determined from daily rainfall totals using the analytical model of *Gash* [1979]. Values for the vegetation parameters in the model (see section 3.2) as well as the ratio of the mean evaporation rate from a saturated canopy ($\bar{E}$, in mm h^{-1}) to the mean rainfall intensity ($\bar{R}$, in mm h^{-1}), were derived from rainfall, throughfall, and stem flow measurements according to the methods of *Gash and Morton* [1978] and *Gash* [1979]. The following parameter values were obtained for the 6- and the 15-year-old forest, respectively: canopy storage (S), 0.8 and 1.2 mm; free throughfall coefficient (p), 0.60 and 0.56; and $\bar{E}/\bar{R}$, 0.15 and 0.12. The trunk storage capacity (S_t) and the fraction of rainfall going to the trunks (p_t) were set at 0.062 and 0.017 mm, respectively, for both forests. A detailed discussion of measured and modeled prehurricane and posthurricane rainfall interception losses (E_i) has been presented by *Waterloo* [1994]. Suffice it to say here that predictions of E_i by the *Gash* [1979] model deviated by −9.9% and +10.3% from values measured in 1990 at Tulasewa and Koromani, respectively. In spite of the difference in forest age and stand density reflected in the values of S and p, modeled total interception losses for the two forests were similar at 382 mm (18.6% of incident rainfall, P) and 371 mm (18.1% of P) for the pines at Tulasewa and Koromani, respectively.

To also quantify evaporation of moisture stored in the litter layer (E_l), an empirical model was developed which uses the

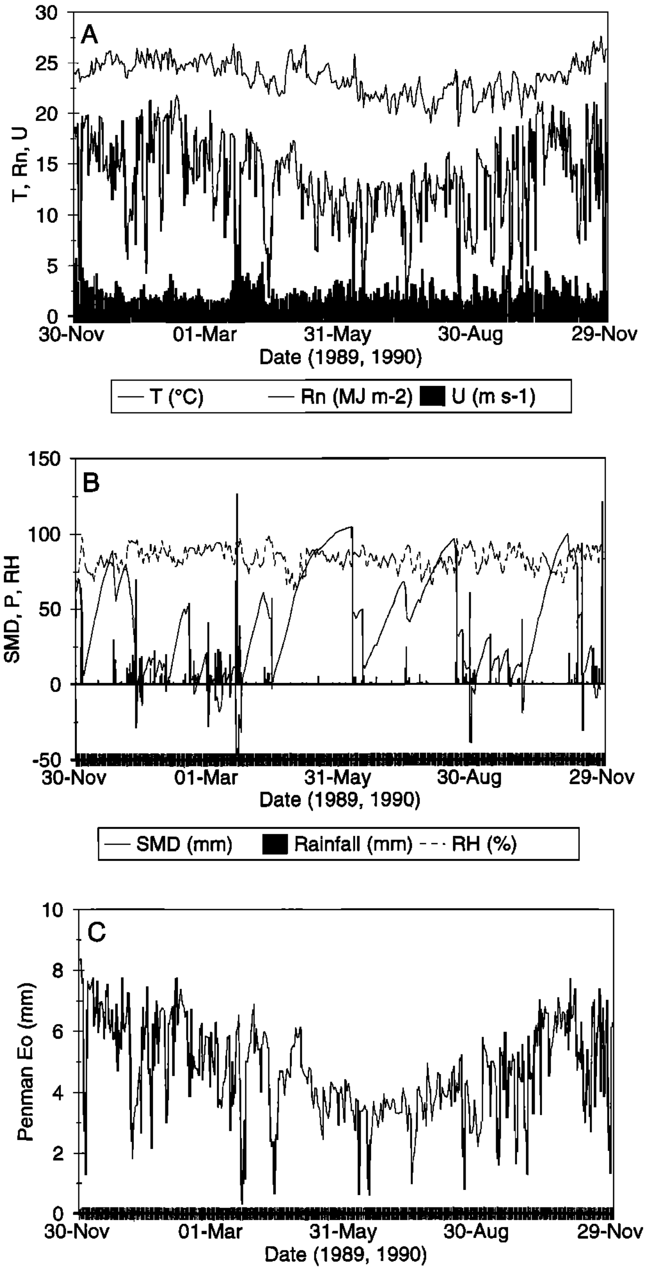


Figure 6. Daily values of (a) net radiation, temperature, windspeed; (b) rainfall, soil moisture deficit, relative humidity; and (c) Penman open water evaporation at Tulasewa forest between November 30, 1989 and November 29, 1990. Missing data have been replaced by data collected at the nearby Nabou station.

daily throughfall totals given by the *Gash* model, a minimum ($S_{l\min}$) and a maximum ($S_{l\max}$) litter layer moisture storage capacity (both expressed in millimeters of water) and an exponential equation (13) describing the decrease in litter layer moisture content (MC_l) due to evaporation as a function of time t (hours) after rainfall according to

$$MC_l = S_{l\max} e^{-0.0135 \cdot t} \tag{14}$$

Evaporation from the litter layer stops when the calculated value of MC_l (equation (14)) falls below $S_{l\min}$. Evaporation from the litter layer during rainfall was assumed negligible. The model was applied using values of 2.4 and 0.2 mm for $S_{l\max}$ and $S_{l\min}$, respectively, for both forests, as derived from gravimetric field measurements of MC_l at various points in time [*Waterloo*, 1994]. The resulting annual totals of E_l were 155 and 153 mm (7.5% of P) for the forests at Tulasewa and Koromani, respectively, bringing the total amounts of rainfall intercepted by the canopy and the litter layer ($E_t + E_l$) at 536 (Tulasewa) and 524 mm (Koromani), or about 26% of P.

4.3.4. Total forest evaporation. To obtain total annual evaporation figures (ET) for the two forests, the values of E_d (which includes E_l) and E_i were summed to give 1926 mm (5.28 mm d^{-1}) for the 6-year-old forest and 1717 mm (4.70 mm d^{-1}) for the 15-year-old stand (Table 2). The bulk of the evaporation consisted of transpiration (E_t, i.e., $E_d - E_l$), which amounted to 72% and 70% of total ET at Tulasewa and Koromani, respectively. The rainfall interception components E_i and E_l made up 20–22% and 8–9% of ET, respectively. The ET/E_0 ratios were high to very high, at 1.13 (Tulasewa) and 1.01 (Koromani). Corresponding ratios of E_t/E_0 were 0.82 and 0.70, respectively (Table 2).

Measured monthly rainfall and predicted amounts of ET, rainfall interception ($E_t + E_l$, Tulasewa forest only) and E_0 for the forests at Tulasewa and Koromani are shown in Figure 7.

5. Discussion

5.1. Radiative and Roughness Characteristics, Aerodynamic and Surface Resistances

The slight but statistically significant difference in shortwave reflection coefficient for the two pine stands (0.100 for the 6-year-old forest and 0.126 for the 15-year-old forest) may be related to differences in stand density and structure [*Jarvis et al.*, 1976]. In addition, the mature forest had lost a considerable proportion of its foliage during the passage of a hurricane in November 1990 [*Waterloo*, 1994], and the lower foliar mass and fresh needles replacing the old ones may have resulted in a higher albedo. The decrease in albedo observed for the stand at Koromani during the dry season of 1991 may presumably be due to an increase in foliar mass during this period [*Waterloo*, 1994]. The values for both stands are within the range given for coniferous forests in the temperate zone (0.08–0.14) [*Jarvis et al.*, 1976]. The albedo for the mature stand is also close to values reported for broadleaved lowland rain forest in Nigeria [*Oguntoyinbo*, 1970] and central Amazonia [*Shuttleworth*, 1988]. However, although the diurnal patterns in albedo for the two pine forests were similar to that observed for the type of tall *Pennisetum polystachyon* grassland they replaced, the overall values for the forest were distinctly lower than the wet season average (0.18) observed for the grassland (Waterloo et al., manuscript in preparation, 1999, Figure 3). This implies that the reflected shortwave component for pine plantations is 28–56% smaller than for grassland, which affects the magnitude of the net radiation considerably. The difference between the expressions relating R_n to R_g at Tulasewa and Koromani (equations (2) and (5)) can therefore at least partly be explained by the observed differences in albedo. Daytime values of R_n were markedly higher for the two forests compared to the grassland site, but the effect was compensated in the case of the Koromani stand where low values of R_n at night (equation (4)) caused overall 24-hour totals of R_n to be similar to those observed for the grassland (Waterloo et al., manuscript in preparation, 1999).

Table 2. Comparison of Water Balance Components at the Tulasewa and Koromani Forest Plots With Those at the Nabou Grassland Plot

	Nabou	Tulasewa	Koromani	Oleolega*
Vegetation	Grass	6-year-old pines	15-year-old pines	15-year-old pines
Stem density	···	825	621	459
Length of record	365	365	365	314
P	2054	2054	2054	1547
E_t	551	1394	1198	···
E_i	93	382	371	···
E_l	102	155	153	···
ET	746	1926	1717	1301
E_0	1701	1701	1701	1396
ET/E_0	0.44	1.13	1.01	0.93
E_t/E_0	0.32	0.82	0.70	···
Q (+ ΔS + L)	1308	128	337	246

Tulasewa and Koromani data from (Waterloo et al., manuscript in preparation, 1999). Values for the Olelolega catchment [*Waterloo*, 1994] added for comparison. All values in millimeters except stem density (stems ha^{-1}) and record length (days). Q, drainage or streamflow; ΔS, change in soil water and groundwater storages; L, catchment leakage.

*Catchment water balance study, January 4 through November 13, 1990.

As for stand roughness characteristics, the ratios of z_0 to h_v derived for the young stand and the mature stand (0.12 and 0.08, respectively) are within the range (0.02–0.14) reported for temperate pine forests by *Jarvis et al.* [1976]. However, the very similar ratios of d to h_v found for the present forests (0.56–0.58) are at the lower end of the spectrum (0.6–0.9) given by *Jarvis et al.* [1976], which may be attributed to the rather open character of the canopies at the study sites as a result of growth stage (Tulasewa) and reduced foliage from past hurricane damage (Koromani). The average daytime course of r_a for the pine stands showed a similar pattern to that observed for the grassland (Figure 4), although at lower values (typically 10 s m^{-1} versus about 25–30 s m^{-1} in the case of the grassland). The difference is to be attributed to the contrast in height and vegetation structure between the grass and the forests and to a much lesser degree to differences in wind speed between sites [*Waterloo*, 1994].

A similar contrast between forest and grassland was observed for the surface resistance against evaporation (Figure 5). The somewhat lower values of r_s observed for the forest at Tulasewa (35–50 s m^{-1}) compared to those at Koromani (60–80 s m^{-1}) reflect the lower LAI value and presumably also the less vigorous growth of the latter.

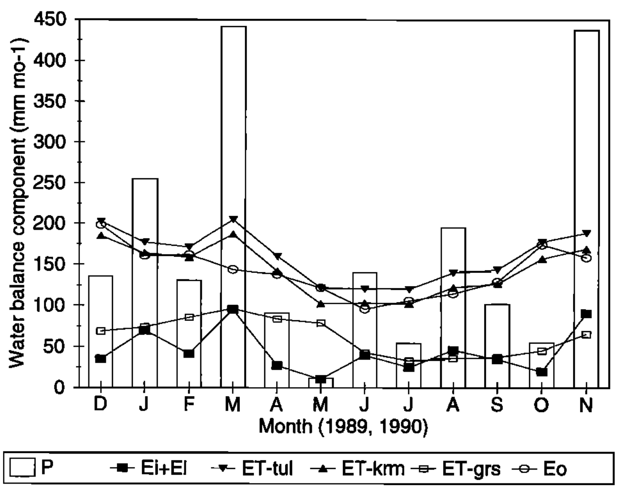


Figure 7. Monthly amounts of rainfall (P), modeled evaporation (ET), and interception losses (E_i + E_l, Tulasewa only) for young (Tulasewa) and mature (Koromani) pine forests. Values of ET for *Pennisetum polystachyon* grassland (ET-grs (Waterloo et al., manuscript in preparation, 1999)) and the Penman open water evaporation (E_0) added for comparison.

5.2. Forest Evaporation

The very high estimates of annual evaporation (ET), transpiration (E_t), and to a lesser extent rainfall interception (E_i) for the two pine plantations of the present study (Table 2) are remarkably similar to the values obtained by *Richardson* [1982] for a 19-year-old stand of *Pinus caribaea* at 700 m asl in Jamaica, namely, 1849, 1365, and 485 mm (15% of P, using a stem flow fraction of 2%), respectively. The Jamaican study comprised a comparison of the water balance of two small but 'predominantly impermeable' catchments, one covered with the cited pine plantation, the other with lower montane rain forest. *Richardson* [1982] ascribed the markedly high total ET to the prevailing high temperatures and breezy conditions. On the other hand, much lower values values for ET and E_t were reported for various upland coniferous plantations and rain forests in Java [*Bruijnzeel*, 1988] and Kenya [*Blackie*, 1979]. This led *Bruijnzeel* [1990] in his review of tropical rain forest water use to consider the Jamaican results to be possibly erroneous because of catchment leakage, something which Richardson (L. A. Bruijnzeel, personal communication, 1998) did not exclude herself. However, the presently found results seem to confirm the findings of *Richardson* [1982], a conclusion which is supported further by the similarly high ET value derived by *Waterloo* [1994] for the pine-forested Oleolega catchment in the study area (Table 2). The difference in annual ET for the 15-year-old pines of the Koromani stand and the ones of the same age at Oleolega almost matches the contrast in stem density (Table 2). As such, the very high evaporation totals obtained by *Richardson* [1982] and the present study seem to be a characteristic of the tropical maritime climate away from the equator.

Thus far, reports of very high evaporation under maritime tropical conditions have been limited to rainfall interception rates [*Bruijnzeel and Wiersum*, 1987; *Scatena*, 1990; *Dykes*, 1997]. However, "direct" measurements of transpiration (as opposed to values derived by subtracting E_i from ET) under

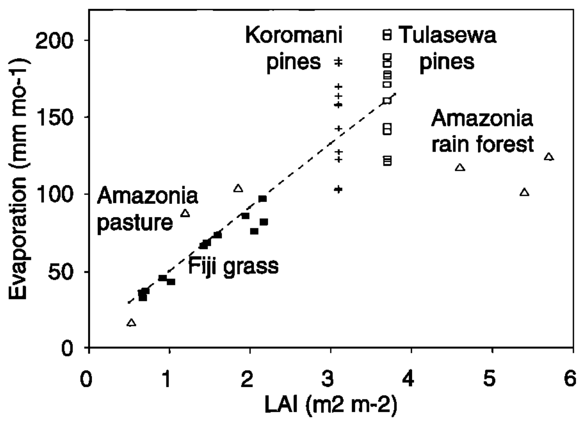


Figure 8. Predicted monthly evaporation totals versus leaf area index for young (Tulasewa) and mature (Koromani) pine plantations and tall *Pennisetum polystachyon* grassland (Waterloo et al., manuscript in preparation, 1999) in southwestern Viti Levu, Fiji. Values for pasture and rain forest in central Amazonia [*Wright et al.*, 1992; *McWilliam et al.*, 1993; *Hodnett et al.*, 1996] are included for comparison.

such conditions are largely lacking [*Bruijnzeel*, 1990] or questionable [*Kline et al.*, 1970]. The most likely source to provide the extra energy that is required to maintain such high rates of evaporation is the ocean. The present study sites were all close to the Pacific Ocean (Figure 1), whose average surface temperature varies between 25° and 28°C [*Basher*, 1986b]. Conversely, the mean daily temperature on days receiving more than 10 mm of rain was 24.2 ± 1.8°C [*Waterloo*, 1994]. As such, a horizontal temperature gradient followed by a downward heat flux is likely to develop whenever the temperature over the land surface drops sufficiently [*Pearce et al.*, 1980]. Further support for this contention comes from the observation of *Waterloo* [1994], that the average rates of evaporation from the wetted pine canopies, as inferred from throughfall and stem flow measurements, were much higher (at 0.72–0.78 mm h^{-1}) than those calculated with the Penman-Monteith equation (0.08–0.30 mm h^{-1}). The fact that forests in the outer tropical region are often subjected to repeated cyclone damage and therefore have a greater roughness than equatorial forests may also play a role in enhancing the (wet) canopy evaporation. The classical approach of determining the surface roughness parameters from above-canopy wind speed profile measurements is perhaps not adequate under these conditions as it may underestimate the surface roughness.

5.3. Implications for Water Yield After Grassland Afforestation

The contrast in energy balance, roughness characteristics, aerodynamic resistance, rooting depth, LAI, surface resistance and rainfall interception (4% of P for grass (Waterloo et al., manuscript in preparation, 1999)) between plantations of *Pinus caribaea* and tall *Pennisetum* grassland in the study area are such that a conversion of the latter to the former is bound to produce profound changes in the water yield from reforested catchments. As illustrated by Table 2, the predicted annual ET for the seasonal grassland (746 mm) is substantially lower than the corresponding values for the two mature but hurricane-damaged stands at Koromani and Oleolega. The contrast is even more pronounced in the case of the vigorously growing 6-year-old plantation at Tulasewa (ET = 1926 mm). As such, taking the data in Table 2 at face value, the conversion of seasonal grassland to evergreen pine plantations in southwestern Viti Levu could result in a maximum decrease in water yield of 1180 mm yr^{-1}. The bulk of this decrease (67–72%) can be attributed to the higher transpiration of the trees, whereas the remainder is the result of increased interception by the pine canopy (24–28%) and litter layer (5%).

As shown in Figure 7, the contrast in water use between the pine plantations and the grassland gradually increases as the dry season progresses. During this time most of the grass dies off, whereas the forests hardly suffer soil water shortage (equations (12) and (13)). As such, the frequently voiced claim that reforestation will restore dry season flows [e.g., *Hardjono*, 1980; *Bartarya*, 1989] is not supported by the present findings. *Bruijnzeel* [1989] argued that reforestation can only be expected to increase dry season water yields if the gain in flow due to the improvement of soil infiltration characteristics that usually accompanies forestation [*Pritchett*, 1979] exceeds the loss of water associated with the higher water use of the forest. The present data confirm that the contrast in water use between the original grassland and the fast-growing pines far exceed the benefits of increased infiltration [*Gilmour et al.*, 1987] under the specific climatic and geographic conditions in Southwest Viti Levu.

Monthly evaporation totals for the two pine plantations and the seasonal grassland are compared as a function of LAI (forest undergrowth LAI not included) in Figure 8. As can be expected on the basis of the information given in the previous sections, the dependency of ET on LAI is much more developed in the case of the grass. Also, the contrast in ET between a *Brachiaria* pasture [*Wright et al.*, 1992] and a lowland rain forest [*Shuttleworth*, 1988] in central Amazonia is much smaller than that between the Fijian grassland and the pine forests, confirming the contention of *Shuttleworth* [1989], that effects of tropical deforestation (and by implication, reforestation) on runoff are likely to be more pronounced for continental edge and island locations than for midcontinental situations, such as central Amazonia.

When considering the effect of grassland reforestation the impact of spatial and temporal variations in rainfall on grassland evaporation in the study area should be taken into account. Although the work in 1990–1991 was conducted during a period for which the rainfall distribution resembled the long-term average at Nabou [*Waterloo*, 1994], the simulations were carried out using climatic data collected at the more inland site of Tulasewa. Rainfall in southwestern Viti Levu shows a high interannual variation (864–2983 mm yr^{-1} at Nadi) and increases with distance from the coast (from 1796 mm near the coast to 2113 mm at Tulasewa in 1990 [*Waterloo et al.*, 1997]). The annual transpiration of the grass may vary considerably with rainfall and the length of the dry and wet seasons [*Riou*, 1984], as these particularly affect the magnitude of the LAI (Figure 8). Rainfall interception losses for the pine plantations will also reflect the variations in rainfall. A related aspect concerns the recurring passage of hurricanes through the study area, which tend to affect the tree plantations more than the grassland. The chief effects include the gradual thinning of the stands and the regular loss of large amounts of foliage, both of which will have a bearing on the magnitude of LAI and therefore (Figure 8) on ET. Consequently, any changes in water yield as a result of grassland reforestation may show considerable spatial and temporal variation.

Finally, it should be recognized that the ET values presented in Table 2 are based on plot measurements that cannot be extrapolated directly to the catchment scale. First, evaporation from grassland catchments in the study area will be higher than the plot-based estimate because of the presence of a riparian vegetation over 10–20% of the catchment. This riparian vegetation generally transpires at the potential rate because it is rarely, if ever, water limited. Second, advection effects will be important, particularly during the dry season, when water use by the dying grass on the surrounding slopes will be minimal and energy partition above the grass will be used mainly to increase the sensible-heat flux (Waterloo et al., manuscript in preparation, 1999). As such, the total ET of the riparian vegetation may well equal or exceed that of the pine forests. "Ballpark" estimates of the ET for a typical grassland catchment would thus be in the order of 860 to 980 mm yr^{-1} for catchments with 10% and 20% riparian vegetation, respectively, assuming the latter would evaporate at the same rate as the young pine stand. On the other hand, evaporation from forested catchments in the study area may be lower than the values obtained here for the forest plots because the latter were relatively well stocked. Catchment-wide tree densities are often much lower, particularly in older stands, because of noncomplete planting, hurricane and fire damage, and mortality on ridges (shallow soils) or in swampy areas. The result obtained over a 10.5-month period for the moderately stocked Oleolega catchment (Table 2) may be used to provide a more realistic estimate of the annual ET for a typical mature pine forest catchment in the area. Combining the ET/E_0 ratio for the Oleolega catchment (0.93) with the annual E_0 for the Tulasewa site yields a value for ET of 1585 mm, that is, 8% lower than the value obtained for the mature forest at Koromani. In summary, the decrease in water yield after reforestation of grassland catchments in southwest Viti Levu is likely to be closer to 500–700 mm yr^{-1}, depending on forest stocking and the size of the riparian zone.

6. Conclusions

This paper is the first to provide a description of the energy partition and water use of plantations of *Pinus caribaea* at two stages of growth in the lowland tropics. Using micrometeorological and hydrological measuring and modelling techniques plausible estimates of both dry canopy evaporation and rainfall interception by the tree canopy and the litter layer have been obtained. The results illustrate the very high rates of evaporation from forests that may be sustained under maritime tropical conditions compared with mid continental locations [*Shuttleworth*, 1989; *Bruijnzeel*, 1990]. As a result, substantial reductions in catchment water yield can be expected, and have been demonstrated, after reforesting the tall seasonal grasslands of the study area with fast-growing pines, particularly during the dry season when much of the above-ground biomass of the grass dies off.

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